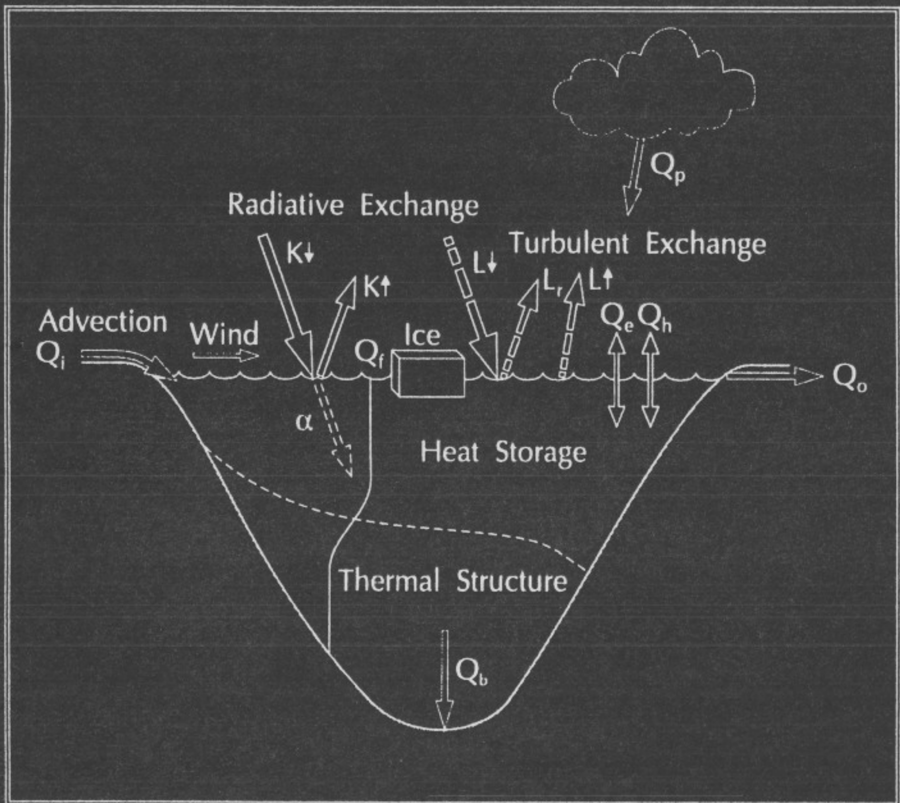
POTENTIAL CLIMATE CHANGE EFFECTS ON GREAT LAKES HYDRODYNAMICS AND WATER QUALITY



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Chapter 2
CLIMATE AND LAKE RESPONSES

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ABSTRACT : *The study of the effects of climate on basin hydrology and lake responses, using examples primarily drawn from the Great Lakes region, is explained in three parts. Firstly, the base case 'current' climate elements are constructed from more than 30 years of records (Sections 2.1 to 2.4). Secondly the climate change case is examined through modeling studies using extreme variations in the observations, steady-state Global Circulation Model (GCM) scenarios and transposition climates (Sections 2.4 to 2.7). When compared to the base case observations, these preliminary Great Lakes results indicate increases in evaporation, runoff reduction, and disruptions to lake thermal stratification characteristics due to climate warming. Thirdly, preliminary results from other lake systems reinforced findings determined from the Great Lakes case (Section 2.8). These results are preliminary and their interpretation requires caution, particularly on the limitations and capabilities of the models used which is discussed in subsequent chapters.*

2.1 INTRODUCTION

Climate has a pronounced influence on large lake hydrodynamics, water quality and ecosystem components. Surface heating and wind mixing affect the seasonal thermal stratification cycle which affects vertical and horizontal circulation patterns. Wind also plays a dominant role in lake set-up, waves and near-shore processes such as up-welling and down-welling. Climate can also influence the seasonal basin and lake hydrology, and consequently, water levels, inflows and outflows and the redistribution of tributary inputs. Consequently, lake hydrodynamic processes are strongly linked to

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climatic and hydrological characteristics over the lake basin. Together, climate, hydrology and lake hydrodynamics can affect the distribution of water quality components within the lake system.

Climatological investigations are largely concerned with heat, moisture and momentum exchanges. Such studies have been conducted on various spatial (global, continental and regional) and temporal (daily, monthly, seasonal, annual and decadal) scales. A climatology can be derived by analysis of meteorological (weather) time series data to quantify the means and variability inherent for a particular area. Climatological (meteorological and hydrological) time-series data are particularly relevant for hydrodynamic and water quality investigations of large lake systems.

Recently, there has been increasing concern that human activities such as fossil fuel burning is progressively altering the chemical composition of the atmosphere. Global atmospheric monitoring has confirmed rapidly increasing greenhouse gas (GHG) concentrations. Part of the concern is that changes in radiative forcing of the climate system may be occurring as a result of anthropogenic loading due to burning of fossil fuels (CO_2, CH_4, N_2O and O_3), changing land uses (CO_2, CH_4) and CFC emissions. General circulation models (GCM's) of the earth's atmosphere have indicated that under scenarios of doubled atmospheric CO_2 concentration (i.e. $2\times CO_2$) the natural greenhouse effect of the atmosphere will be enhanced. Such enhancement is expected to affect the climate system resulting in higher global atmospheric temperatures and other climate variables. Changes in the climatological characteristics are expected to vary over the globe. Climate warming has the potential to affect the physical, chemical and biological characteristics of a region including aquatic systems.

Investigation of the current climatological conditions and potential changes on global and regional scales due to projected climate warming is ongoing. Several global conferences have been held to assess probable impacts of climate warming on various sectors of the environment and economy in various nations and to assess potential adaptive/mitigative strategies (IPCC 1990, 1992). From the perspective of aquatic systems, a significant amount of research has been conducted on the North American Great Lakes. The Great Lakes are a dominant feature on the North American Continent representing the largest continuous volume of freshwater on Earth. Decades of research have been undertaken to establish the physical, chemical and biological characteristics of this system, including lake hydrodynamics and water quality. Combined monitoring and research conducted in Canada and USA has resulted in a wealth of environmental data to establish climatologies and for assessing probable impact of climatic changes on large lake systems. Such results have implications and applications to other large lake systems over the world. While considering research conducted elsewhere, the focus of this Chapter is directed primarily on the climatological, hydrological aspects related to lake hydrodynamics and water quality in the Great Lakes.

This Chapter focuses on climate and provides a discussion of dynamic responses of lakes forced by current climate and potential climatic changes. As background to this and succeeding Chapters concerned with the Laurentian Great Lakes, physical characteristics and climatic measurement networks are briefly described. Selected key weather elements of the Laurentian Great Lakes are described to provide a basis of the current temporal and spatial climatic variability which have an effect on lake hydrodynamics and water quality. From a regional perspective, monitoring and modeling of basin and Great Lakes hydrology and energy balance are described which supports discussions on vertical mixing, large-scale circulation, waves, and water quality discussed in later Chapters. Important climate warming concerns are addressed by providing a discussion on climate change and regional scenarios with a synopsis of case study examples which attempt to provide insight into potential changes to climatological characteristics (hydrological and lake heat budget) which impact on lake hydrodynamics and water quality. The final section of this chapter will show that the climate change research conducted over climatic regimes in other parts of the world reinforces many of the potential impacts on hydrothermal dynamics from Great Lakes investigations.

2.2 GREAT LAKES PHYSICAL CHARACTERISTICS

The Great Lakes Basin, shown in Figure 2-1a, contains an area of approximately 770,000 km^2, about one-third of which is water surface (Freeman and Haras, 1978). The Great Lakes basin resides in Canada (44%) and the USA (56%). The basin extends some 3,200 km from the western edge of Lake Superior to the Moses-Saunders Power Dam on the St. Lawrence River. The water surface drops in a cascade over this distance some 180 m to sea level. A general profile of the Great Lakes (Figure 2-1b) shows that Lakes Huron, Michigan and Ontario have comparable depths, that Lake Erie is the shallowest and that Lake Superior is the deepest and largest. Table 2-1 contains pertinent gross statistics on the sizes of the Great Lakes, Lake St. Clair, and their basins. Table 2-2 provides long-term mean flow statistics for Great Lakes connecting channels and diversions.

Lake Superior has two inter-basin diversions of water into the system from the Hudson Bay Basin: the Long Lac and Ogoki Diversions. Lake Superior waters flow through the lock and compensating works at Sault St. Marie and down the St. Marys River into Lake Huron where it is joined by water flowing from Lake Michigan. Lake Superior is completely regulated, to balance Lakes Superior, Michigan, and Huron water levels. Lakes Michigan and Huron are considered to be one lake hydraulically because of their connection through the deep Straits of Mackinac. The second inter-basin diversion takes place from Lake Michigan at Chicago. Here water is diverted from the Great Lakes to the Mississippi River Basin. The water flows from Lake Huron through the St. Clair River, Lake St. Clair, and Detroit River system into Lake Erie. The drop in water surface between Lakes Michigan-Huron and Lake Erie is only about 2 m. This results in a large backwater effect between Lakes Erie, St. Clair, and Michigan-Huron; changes in Lakes St. Clair and Erie levels are transmitted upstream to Lakes Michigan and Huron. From Lake Erie, the flow is through the Niagara River and Welland Diversion into Lake Ontario. The major drop over Niagara Falls precludes changes on Lake Ontario from being transmitted to the upstream lakes. The Welland Diversion is an intra-basin diversion bypassing Niagara Falls and is used for navigation and hydropower. There is also a small diversion into the New York State Barge Canal System which is ultimately discharged into Lake Ontario. Lake Ontario is completely

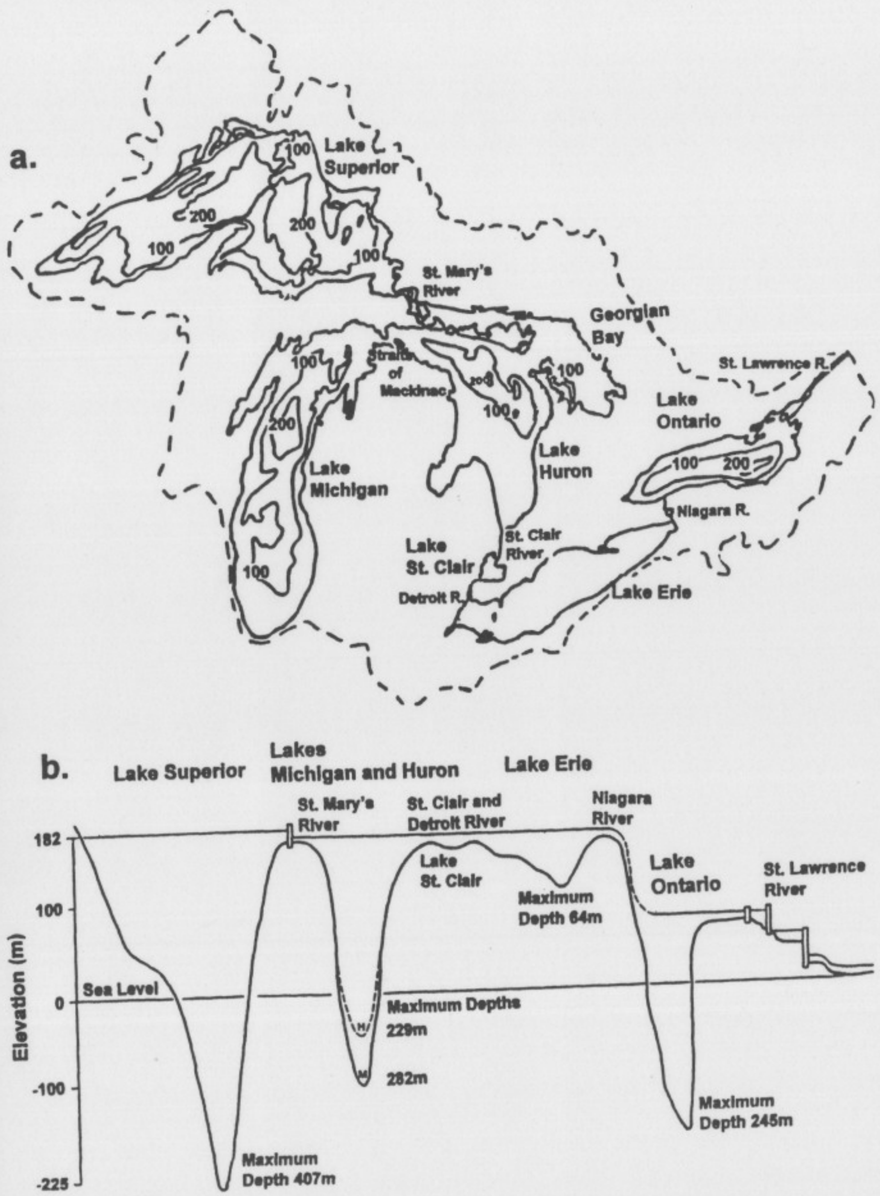

Figure 2-1. *The Great Lakes basin depicting (a) the bathymetry of the Great Lakes, and (b) a profile of the Great Lakes (Rodgers, 1969).*

		SUP.	MIC.	HUR.	ERI.	ONT.
Low Water Datum (LWD)	(m)	182.9	175.8	175.8	173.3	74.0
Length	(km)	563	494	331	388	311
Width	(km)	259	190	294	92	85
Shoreline Length	(km)	4,795	2,670	5,120	1,377	1,168
Total Surface Area	(km^2)	82,100	57,750	59,500	25,320	19,000
Volume at LWD	(km^3)	12,230	4,920	3,537	470	1,637
Mean Depth < LWD	(m)	149	85	59	18.7	86
Max. Depth (LWD)	(m)	407	282	229	64	245
Mean Surface Level(IGLD)	(m)	183.1	176.5	176.5	74.7	74.7

Table 2-1. *Physical characteristics of the Great Lakes [Based on Upchurch (1976)]*

	Mean Monthly Discharge (m^3/s)	Maximum Monthly Discharge (m^3/s)	Minimum Monthly Discharge (m^3/s)
	Connecting Channels		
St. Marys River	2,140	3740	1160
Straits of Mackinaw*	1,500	-	-
St. Clair River	5,180	6740	3000
Detroit River	5,320	7080	3170
Niagara River	5,740	7620	3330
St. Lawrence River	6,910	10100	4360
	Diversions		
Ogoki River & Long Lake	111	329	0
Long Lac Diversion	41	134	0
Chicago Diversion	141	323	49
Welland Canal	146	286	8
New York State Canal System *	20	-	-

Table 2-2. *Long-term mean and range of Great Lakes flows through connecting channels and diversions [David Fay, Personal Communication, Coordinating Committee on Great Lakes Basic Hydraulic and Hydrological data 1996; * Upchurch (1976)].*

regulated to balance damages upstream on Lake Ontario with those downstream on the St. Lawrence River. The outflows are controlled by the Moses-Saunders Power Dam between Massena, New York and Cornwall, Ontario. From Lake Ontario, the water flows through the St. Lawrence River to the Gulf of St. Lawrence and to the ocean.

2.3 CLIMATIC MEASUREMENT NETWORKS

2.3.1 Land observations

Meteorological, hydrological, and limnological data are observed, processed and archived by several organizations within both countries and are readily accessible for research on Great Lakes large-scale or site specific analyses. Standard meteorological observations are made hourly (also 3 and 8 hourly) at synoptic stations, augmented by automatic stations and volunteer observation sites. There are approximately 2,000 climate (temperature and precipitation) stations in the Great Lakes Basin (Croley et al., 1996). Standard observations include key variables such as air temperature, relative humidity, wind speed and direction, sunshine, visibility, pressure and cloudiness etc. Figures 2-2a and 2-2b provides an indication of the spatial distribution of stations observing key meteorological data over the basin as well as an indication of the coverage at the periphery of the Great Lakes. Long-term data records are available from the Atmospheric Environment Service (AES) in Canada and the National Climate Data Centre (NCDC) in the USA. Hydrological observations (stream discharge, groundwater etc.) are conducted by the Water Survey of Canada (WSC) and the US Geological Survey (USGS) in the USA. Hydrological measurements are not generally continuous and as such these data are analyzed statistically to provide flow and discharge estimates. Estimation of the flows at connecting channels are an important consideration in lake hydrological investigations. Coordinated flow and water level data are provided to the International Joint Commission (Coordinating Committee, 1977).

2.3.2 Lake observations

In situ observations are conducted by fixed moorings (meteorological buoys) which observe standard variables such as air temperature, relative humidity, wind speed and direction. Standard buoys can also be deployed to measure in situ radiation components (i.e. incoming global solar radiation, incoming long-wave radiation and net radiation). In conjunction with the meteorological buoy observations, additional specialized observations including water temperature and current speed and direction are observed at specific depths. Buoy observations are normally deployed in the ice free season and have a sampling frequency on the order of 10 to 20 minutes depending on storage capability. On large lakes such as the Great Lakes, in situ observations are often used for specialized studies. For whole lake observations, ship surveillance is conducted over a grid of stations in surveys generally lasting a week in duration. A ship offers the flexibility for detailed observations at a range of sites over the lake, however, in any particular year whole lake surveys are conducted only say 3-8 times per year in the lower Great Lakes and infrequently in the upper Great Lakes. Measurement of surface water temperature have been conducted by airborne radiometer technique and more recently satellite observations have been implemented.

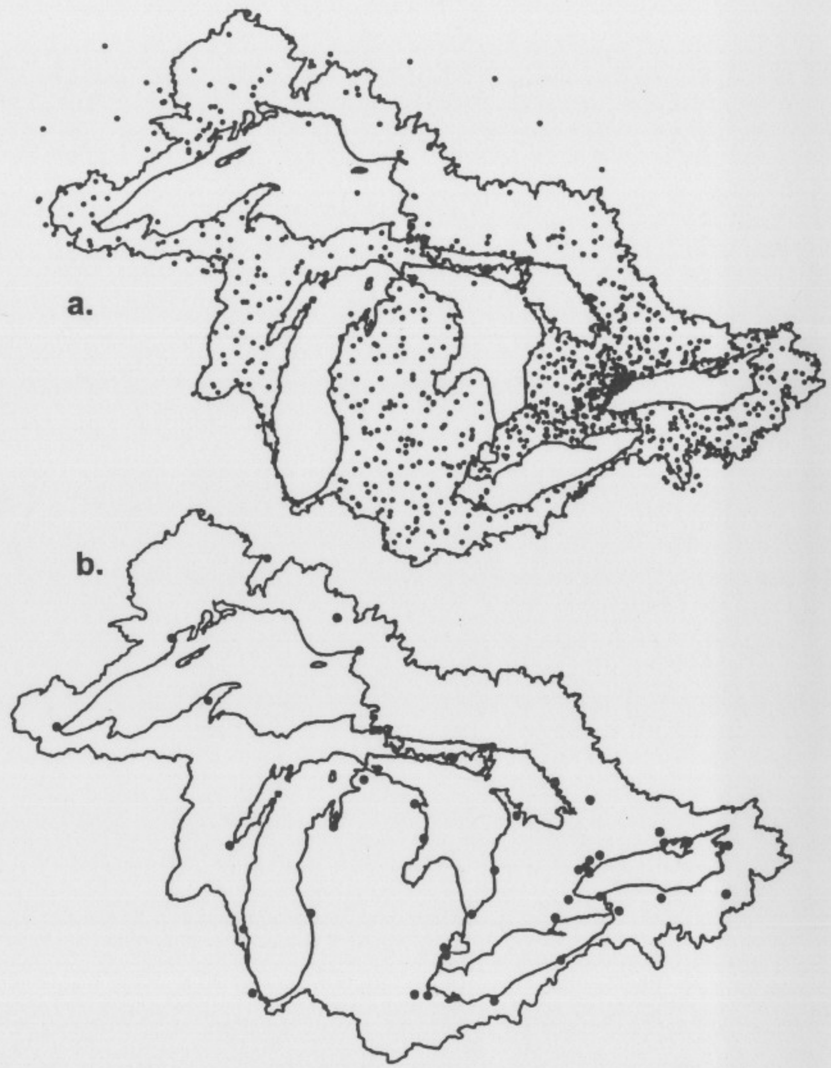

Figure 2-2. *Meteorological station distribution within the Great Lakes Basin depicting (a) temperature and precipitation stations, and (b) temperature, humidity wind speed and cloud cover stations (Croley et al., 1996).*

2.3.3 Over-lake meteorological fields

As indicated in Figures 2-2a and 2-2b, the land station network around the Great Lakes basin and at the perimeter of the lakes is generally very dense. However, the lakes are largely devoid of permanent climatological sampling. Due to the moderating influence of the Great Lakes, spatial and temporal differences occur between lake and land observations. Consequently, specification of meteorological fields over the large expanse of the Great Lakes is one of the largest difficulties encountered in computing heat, moisture, and momentum exchanges as well as lake circulation patterns.

Adjustment of land station data (wind, temperature, and humidity) to over-lake estimates have been accomplished either through ratios or more complex regression techniques (Phillips and Irbe, 1978; Resio and Vincent, 1977). Based on detailed International Field Year for the Great Lakes (IFYGL) experiments, adjustments have been developed to consider the effect of atmospheric stability, fetch, wind speed class and duration over water. Phillips and Irbe (1978) summarized comparisons between simultaneous lake and land measurements and the development of ratios and regression equations for application to Lake Ontario. Modern computations of lake energy balance incorporate corrections to land-based observations. Additional corrections include modification of wind, temperature, and humidity to standard observation heights.

2.4 GREAT LAKES BASIN CLIMATIC CHARACTERISTICS

The climate of the Great Lakes basin is largely influenced by latitude, its continental location, large-scale circulation patterns and the lakes themselves. Climatic conditions and weather have a pronounced effect on lake hydrodynamics and can significantly affect water quality conditions especially in susceptible lakes and embayments. The climatic conditions can be quite varied over the large expanse of the Great Lakes basin. The following contains a general description of the selected key meteorological variables which affect hydrological, limnological and water quality conditions of the Great Lakes region.

2.4.1 Air mass circulation and storms

Continental climate conditions dominate the basin while the lakes moderate the climate toward semi-marine conditions (Phillips and McCulloch, 1972). The basin is influenced by the warm moist air of the Pacific and the Gulf of Mexico and cold dry Arctic air masses. During the winter months, cold Arctic air masses cover the central part of the basin approximately 25% of the time, while moist cloudy air masses of Pacific origin dominate the region about 75% of the time. In summer, the northern part of the basin is dominated by cool Pacific air 30-40% of the time but similar air masses occur in the south only about 10% of the time; the frequency of occurrence of hot humid air from the Gulf of Mexico is about 40% in the southern part of the basin and only 10% north of Lake Superior (Sanderson, 1980).

The basin is frequently affected by weather systems which develop along low pressure storm tracks resulting in rather large day-to-day variability in weather conditions. In spring and summer, the basin is subject to individual and squall-line thunderstorms, with about 20 thunderstorm days per year in the north and about 30 such days in the south (Kunkel et al., 1993). These convective storms can produce high rates of precipitation of 70 to 100 mm hr^{-1}, strong gusty winds often to 100 km hr^{-1} and hail. Occasionally they spawn tornadoes, more often in the southern regions of the basin. Severe winter storms with 15 cm or more of snow or glaze can affect the basin from late October through April. These cyclones, usually passing from west to east, can bring heavy snow, sleet, and freezing precipitation to an area 100 to 200 km wide, primarily on the north side of the cyclone trajectory. Winds greater than 50 km hr^{-1} can occur. Such storms can occur for a 12-24 hr duration (Kunkel et al., 1993).

The lakes have a high heat storage capacity due to their large volume. This heat storage and air/water exchange of heat results in modification of the climate around the lake. During spring and summer, the lakes tend to cool the surrounding land mass and in fall and winter they tend to warm them, often resulting in lake-effect storms on the lee of the lakes. The large heat storage capacity is also responsible for ice free periods of most of the deep lakes.

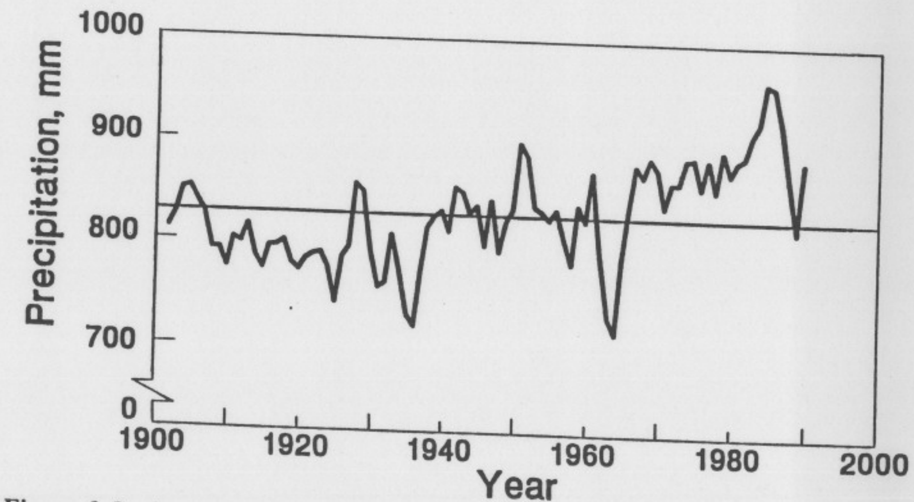

Figure 2-3. *Long-term annual mean precipitation over Lakes Michigan, Huron, St. Clair and Erie for the period 1900 - 1990. (Croley et al., 1996)*

2.4.2 Precipitation

Precipitation is responsible for major long-term variations in lake levels (Quinn and Croley, 1981; Quinn, 1985). Average precipitation over the period 1900 to 1990 was 79cm, 84cm, 84cm, 89cm, 88cm for Lakes Superior, Michigan, Huron, Erie, and Ontario respectively (Croley, 1995). Total annual precipitation over Lakes Michigan-Huron, St. Clair, and Erie for the 1900-90 period is illustrated in Figure 2-3 (Quinn, 1981; Quinn and Norton, 1982). Variability in annual precipitation amounts is quite evident. In general, the majority of years from 1900-1940 are below the mean while 1940 to the present has experienced higher than average precipitation. Extreme lows and high precipitation years are evident. The highest recorded precipitation occurred in 1985 and was 7-33% higher than the 1900-90 average.

Mean annual precipitation distribution over the Great Lakes basin (Figure 2-4a) and seasonal distributions for selected stations (Figure 2-4b) are based on long-term 1951-1980 data (Kunkel et al., 1993). South and east locations of the basin are generally wetter while drier conditions are evident in the north and western portions. As a result of lake-effect precipitation events usually in the fall and early winter, the south and east shores of the lakes tend to higher precipitation accumulations.

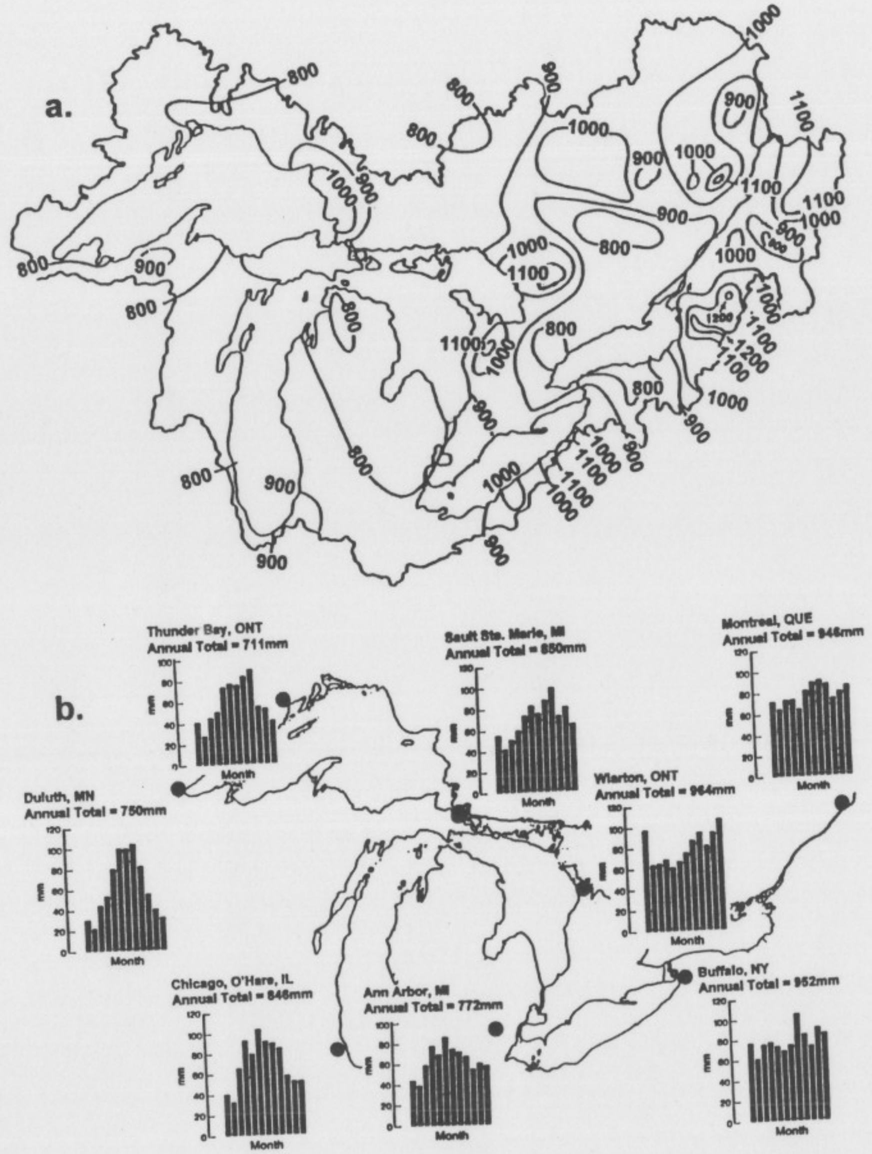

Figure 2-4. *(a) Annual precipitation distribution over the Great Lakes and (b) seasonal variation for selected stations (redrafted from IJC, 1993).*

2.4.3 Air temperature

Long-term annual mean air temperature from stations at the perimeter of the Great Lakes (Figure 2-5) indicate three distinct temperature regimes: a low temperature regime from 1900-1929, a higher temperature regime from about 1930-1959, and an additional low regime from 1960-present period. The difference between the previous and current regime is a drop of about 0.5°C (Croley, 1995).

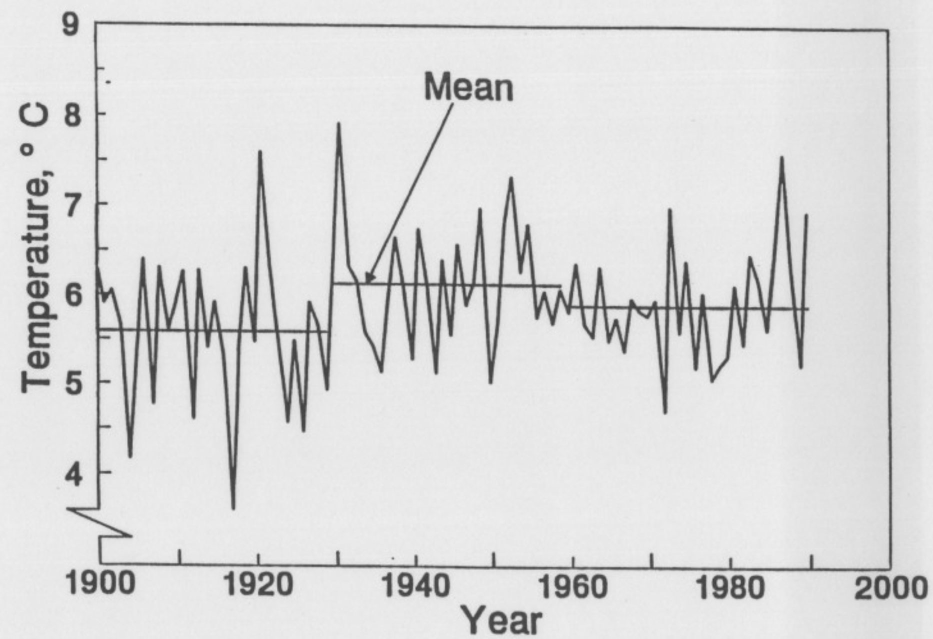

Figure 2-5. *Long-term annual mean air temperature based on stations at the periphery of the Great Lakes (Croley et al., 1996).*

Contours of mean annual temperature (Figure 2-6) show a north south gradient. Mean values range from about 0°C in the north to 10°C in the south. Large seasonal variations in temperature are observed at all stations in the basin and are related to the continental location of the basin as well as its large north-south and east-west extents. In general, stations in the northern and western parts of the basin exhibit a larger seasonal temperature range than do stations to the south and east (Kunkel et al., 1993). Stations in close proximity to the lakes have a smaller annual temperature range than those removed from the moderating influence of the lake which experience both cooler winter and warmer summer temperatures.

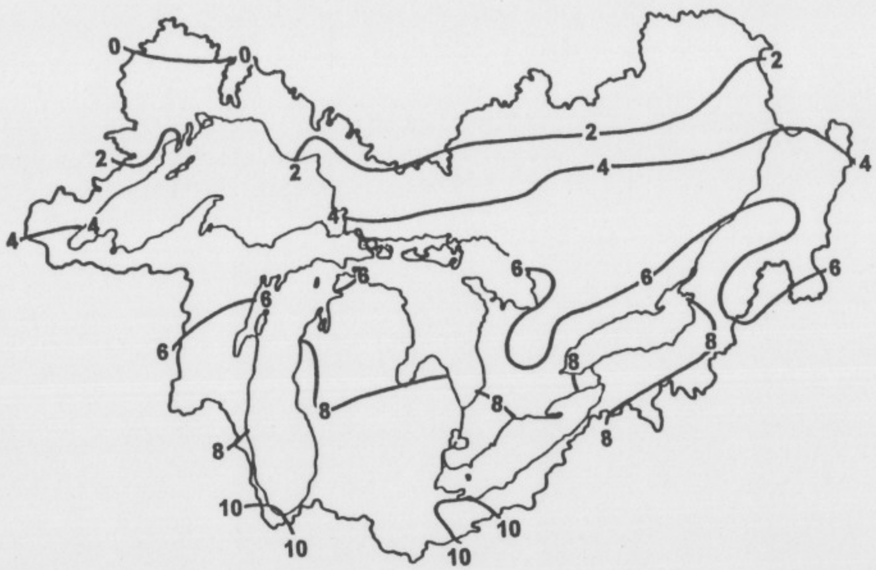

Figure 2-6. *Mean annual air temperature contours over the Great Lakes basin (redrafted from IJC, 1993).*

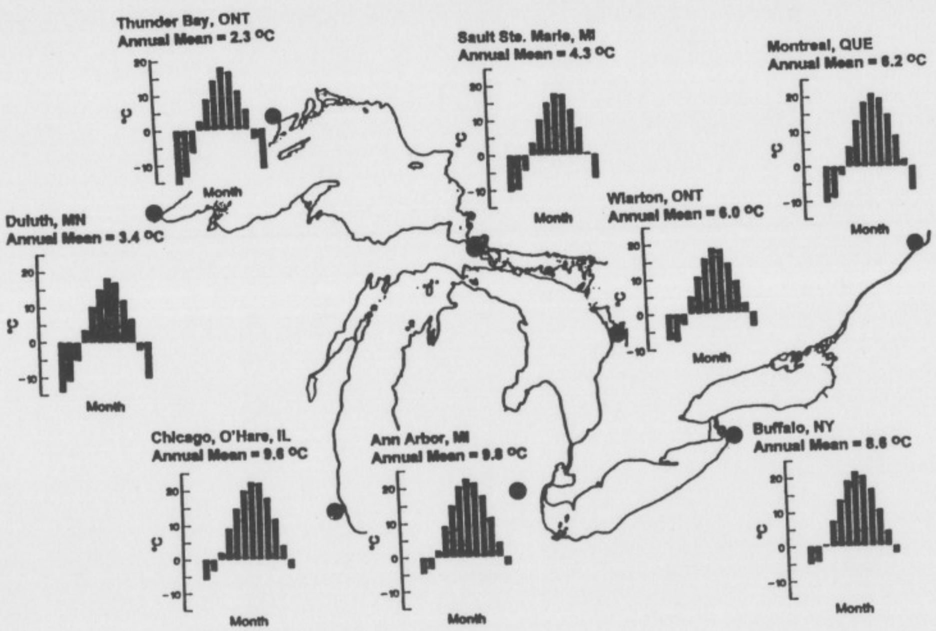

Figure 2-7. *Seasonal Mean and range of air temperature for selected stations around the Great Lakes (redrafted from IJC, 1993).*

Figure 2-7 illustrates the seasonal mean and range of air temperature for selected stations over the Great Lakes basin. Compared to stations located in the southern and eastern parts of the basin, stations to the north and west have a larger seasonal temperature range. Stations in close proximity to the large lakes have a smaller annual temperature range than stations located some distance from the lake (Kunkel et al., 1993). Stations further removed from the lakes also exhibit a larger annual range with cooler winter and warmer summer temperatures compared to those stations in close proximity to the lakes at similar latitudes. The moderating influence of the Great Lakes is related to the large heat storage capacity of the Great Lakes.

2.4.4 Water temperature

Monitoring of water surface temperature on large water bodies the size of the Great Lakes is difficult and expensive. Many techniques have been employed including the use of water intake temperatures, meteorological buoys, lake-wide and basin-wide surveillance, airborne radiometer over flights and satellite measurements. Water intake temperatures are limited to the near-shore local area and meteorological buoy observations are generally used in specialized studies of short duration. Ship surveillance is expensive, however, detailed whole lake observations can be reliably conducted. Difficulties encountered include aliasing of the data especially if the surveys are longer than a week during periods of rapidly changing temperatures. Surveys are conducted primarily in ice free periods and the number of surveys are generally less than 8 per year, being more frequent on the lower Great Lakes. Airborne radiometer over-flights, superseded by satellite observations can be used to augment whole lake ship surveillance.

The latitudinal variation and dimensions of the Great Lakes can significantly affect the seasonal mean and range of temperature and also temporal lags in heating and cooling cycles. Mean surface water temperatures for the Great Lakes are illustrated in Figure 2-8.

A detailed example of the seasonal mean and variation of observed water temperature is shown in Figures 2-9a to 2-9d for the lower Great Lakes. Surface temperatures for Lake Erie (Figure 2-9a) and Lake Ontario(Figure 2-9b) are based on ship surveillance, airborne radiometer over-flights and satellite data over the period 1966-1984 (Schertzer and Sawchuk, 1985). Due to extensive ice cover on shallow Lake Erie, few surveillance observations have been conducted during the winter months. The composite plots indicate that the range of temperature is generally larger for Lake Erie than for the deeper Lake Ontario. Due to the larger volume and heat storage capacity of Lake Ontario, wintertime temperatures are higher compared to Lake Erie. Based on these data, the variation in temperature during the heating and cooling seasons is larger for Lake Ontario and the range in temperature at the summer maximum appears to be larger for Lake Ontario. Figure 2-9c illustrates that the lake-wide mean temperature of Lake Ontario is significantly lower than that of shallow Lake Erie during the summer months, however, the higher heat storage of Lake Ontario results in a higher lake-wide temperature during winter months. Climate (heating and wind mixing etc.) affects parts of a lake differently and this is especially apparent in lakes with large differences in bathymetric characteristics as shown in Figure 2-9d for Lake Erie west (10m depth), central (25m depth) and east (64m depth) basins . The lake-wide mean temperature of Lake Erie closely corresponds to the temperature of the central basin which is the largest portion of the lake. In comparison, the very shallow west basin warms to a higher temperature earlier and cools earlier compared to either of the other basins. The deeper east basin achieves a significantly lower surface water temperature compared to the remaining basins. Lake Erie in particular displays a complex pattern of thermal and hydrodynamic characteristics largely due to the differing bathymetry between basins.

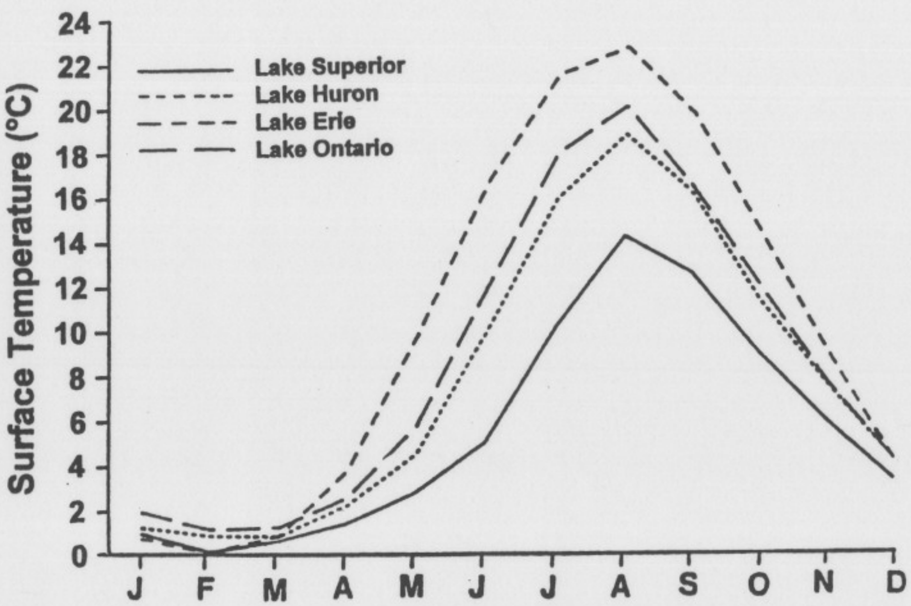

Figure 2-8. *Long-term seasonal mean and range of observed water surface temperature for the Great Lakes (Murthy and Schertzer, 1994).*

Water temperatures during winter are not observed as frequently compared to other times of the year, especially under ice covered conditions. Lake Erie, in particular, develops a significant ice cover in most years (see section 2.4.8 on Ice Cover) and as such water temperature and heat storage estimates during the winter are scarce. The vertical distribution of water temperature during ice cover conditions on Lake Erie was measured by Stewart (1973) by helicopter. In winter, a reverse but weak stratification may occur because water at 4°C is at maximum density and sinks to the lake bottom. Stewart (1973) suggested that the lake was isothermal or nearly isothermal at 0.1°C or less from mid February to mid-March. In mild winters, the open-water areas are exposed to wind action which keeps the water column fully mixed in contrast to the weakly stratified condition which develops under complete ice cover (Schertzer, 1987).

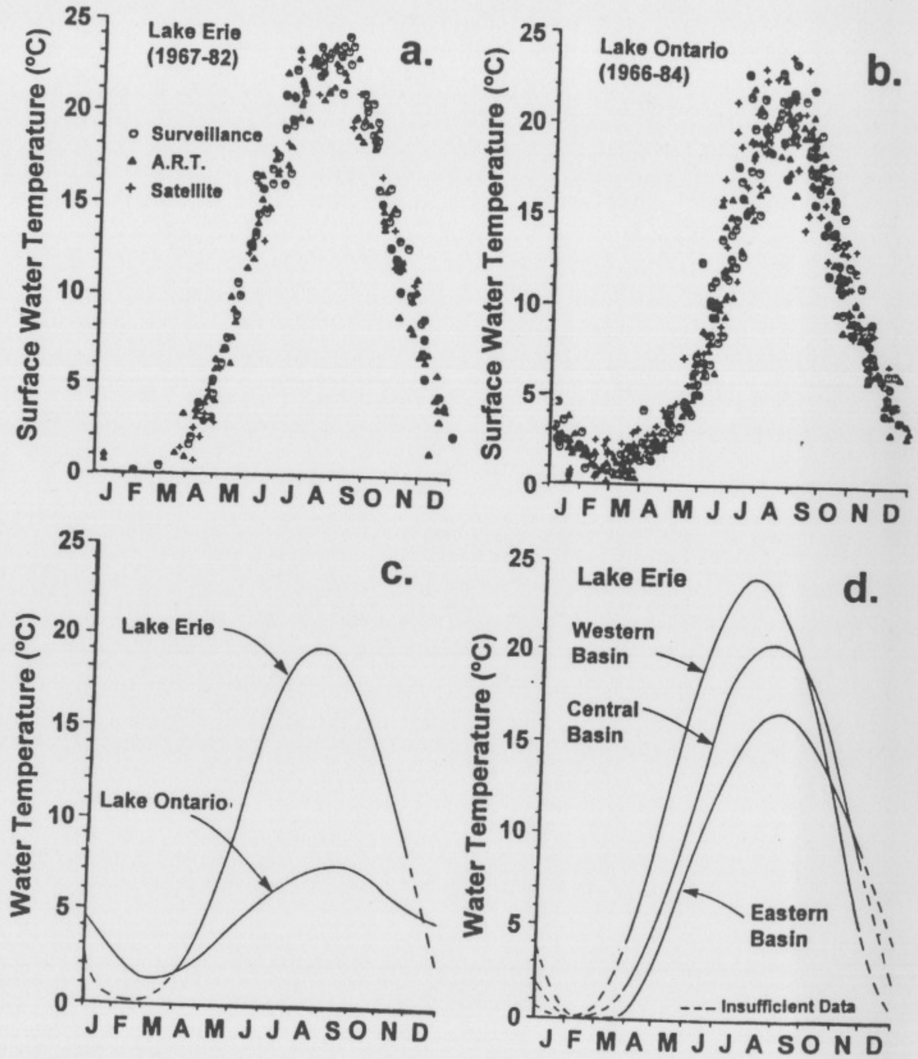

Figure 2-9. *Long-term water temperature observations for the Lower Great Lakes, (a) Lake Erie composite surface temperature, (b) Lake Ontario composite surface temperature, (c) Lake Erie and Lake Ontario lake-wide mean temperature, and (d) comparison of basin temperatures in Lake Erie. (based on Schertzer and Sawchuk, 1985; Schertzer, 1997)*

2.4.5 Winds

Winds constitute one of the principal forces driving lake circulation. Investigations of currents on the Great Lakes rely heavily on wind speed and direction, determined either from over-lake meteorological buoys or from stations at the periphery of the lake.

In terms of modeling lake circulation, lake exposure and completeness of wind records are of prime concern.

Phillips and McCulloch (1972) describe average wind speed and direction for selected stations over the Great Lakes basin and Saulesleja (1986) summarized seasonal mean wind speeds and vector mean wind velocities (see Figures 2-10a to 2-10d).

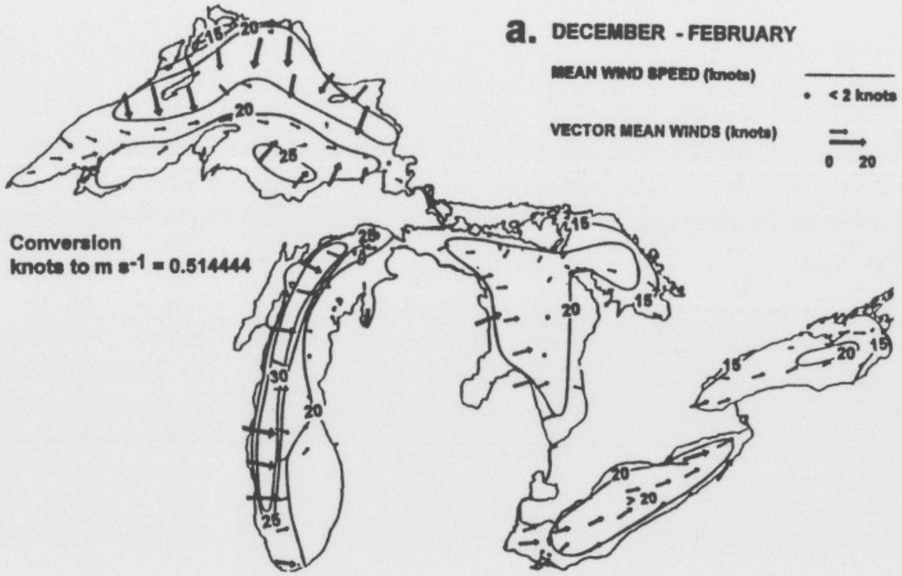

Figure 2-10a. *Seasonal mean wind speed contours and vector mean wind velocity, (a) December-February. (Redrafted from Saulesleja, 1986). Arrows point in the direction of the vector mean with length proportional to its magnitude.*

During winter (Figure 2-10a), wind speeds averaged 2.7-8.5 m s^{-1}. Over the middle and upper lakes region, winds blow from the west and northwest 40-50% of the time with northwest winds prevailing. South of the lakes, winds from the west and southwest predominate 30-40% of the time.

Mean wind speeds during spring (Figure 2-10b) tend to exceed 3.6 m s^{-1} with the highest speeds in excess of 5.8 m s^{-1} associated with cyclonic activity. Southwest winds are more frequent across the lower lakes and northwest winds prevail at higher latitudes.

Summer winds (Figure 2-10c) are generally more variable in direction than winter winds, but less variable in speed. Lake breeze circulations are common along the lake periphery with a frequency of 35% penetrating up to 40 km inland.

The transition between summer and winter winds is depicted in Figure 2-10d for October. During these months, the increase in cyclonic activity and the large thermal differences between air and water contribute to high mean wind speeds. The highest mean wind speeds over the basin exceed 5 m s^{-1} and no mean wind speeds were less than 2.7 m s^{-1} from any direction (Phillips and McCulloch, 1972).

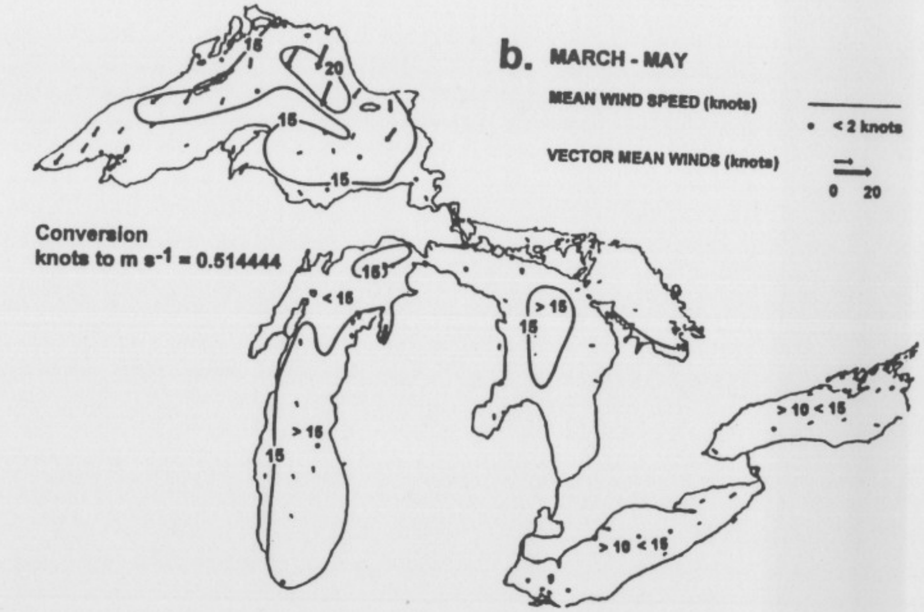

Figure 2-10b. *Continued ... March-May*

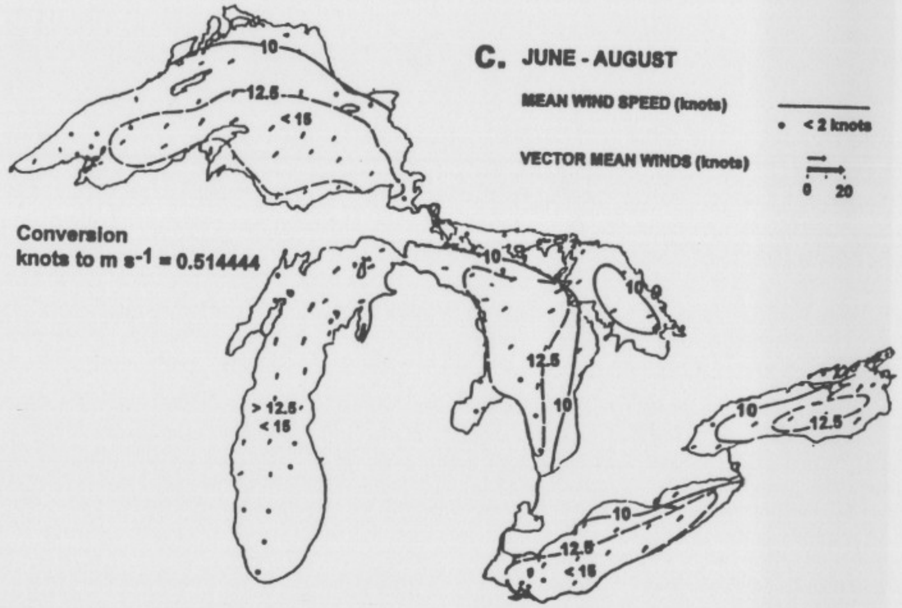

Figure 2-10c. *Continued... June-August.*

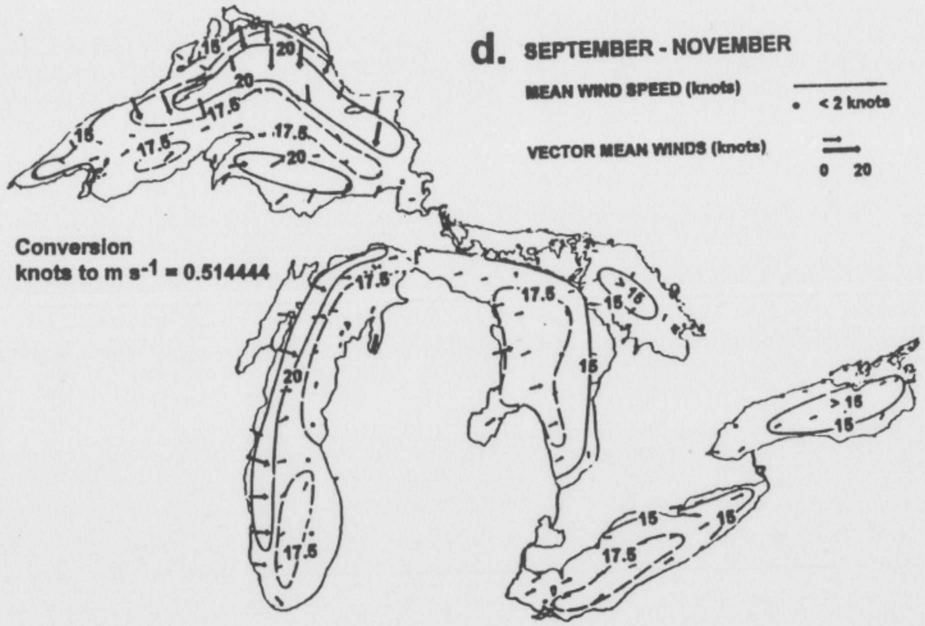

Figure 2-10d. *Continued ... September-November.*

For hydrodynamic evaluations on large lakes, lake exposure is an important consideration. As part of hydrodynamic studies on Lake Ontario (Simons and Schertzer, 1985; Schertzer and Simons, 1985), the effect of site exposure was assessed by comparing long-term wind observations from an inland site (Toronto International Airport) compared to a lake-exposed site (Toronto Island Airport) for the period November to April 1973-1983. Lake exposed wind speeds had significantly higher variance than the inland site with winds averaging 2.7 m s^{-1} higher for easterly winds and 2.0 m s^{-1} higher for southeasterly winds (Figure 2-11). Time-series analysis of the wind stress showed that the along-shore component of the wind stress was greater than the cross-lake component with the computed along-shore wind stress generally larger at the more exposed Toronto Island Airport location. Spectral analyses indicated that there was high coherence between the two stations for periods less than 30 days.

Since the scale of atmospheric weather patterns is typically much larger than an individual Great Lake, the wind field may be expected to be rather uniform over the lake. This hypothesis was tested in Lake Ontario (Simons and Schertzer, 1985) by comparing observed hourly values of land and over-lake station wind speed (> 3 m s^{-1}) and direction with values at a central meteorological buoy (M5) station (Figure 2-12a). Figure 2-12a shows land station locations (i.e. Toronto Airport M1, Toronto Is. M2, Port Hope M4, Point Petre M8, and Point Breeze M7) along with lake stations M3, M5, M6 and M9. The study showed that wind direction tends to deviate left in the western basin of Lake Ontario and right in the eastern basin (Figure 2-12b). The Toronto Island station showed a systematic clockwise deviation of wind direction (about 50 degrees) from an adjacent beach station while Port Hope consistently underestimated the wind speed compared to the mid-lake station.

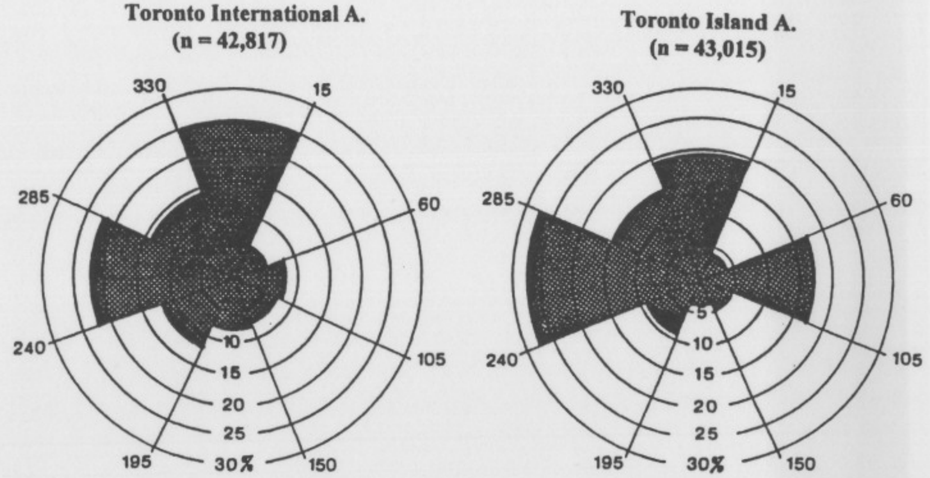

Figure 2-11. *Comparison of wind direction frequency between an inland site (Toronto International Airport) and an lake exposed site (Toronto Island Airport) for Nov. to April 1973-83 for winds greater than 3 m/s (Schertzer and Simons, 1985).*

Power spectra and cross-spectra of wind stress (Figure 2-13a) were computed by the lagged covariance method with maximum lag of 23.4 days. Station rotary coefficients and ellipse orientation (Figure 2-13b) were computed as well as station comparisons through coherence and amplitude ratio (Figure 2-13c). Figures 2-13a to 2-13c show results for selected stations (Toronto Airport M1, M5 and Point Petre M8). The power spectra showed a gradual decrease of energy density with increasing frequency except for a broad peak at periods of about a week. Based on spectral results of all stations, the wind energy is fairly uniform over the lake except for the Port Hope station M4 (Simons and Schertzer, 1985). Rotary spectra of the wind indicated a clockwise shift in prevailing wind directions between the western and eastern ends of Lake Ontario. Such a curvature in the wind field is consistent with the center of atmospheric pressure systems being located to the south of Lake Ontario. Coherence analyses indicated high coherence between over-water stations including Toronto Island with lowest coherence between Port Hope and the mid-lake station at M5. Amplitude ratios of the wind stresses as a function of frequency clearly showed the uniformity of wind speed over the whole lake, the exception being the drastic reduction of wind speed at Port Hope M4 and the substantial reduction at Toronto Beach station M2 (see Simons and Schertzer, 1985).

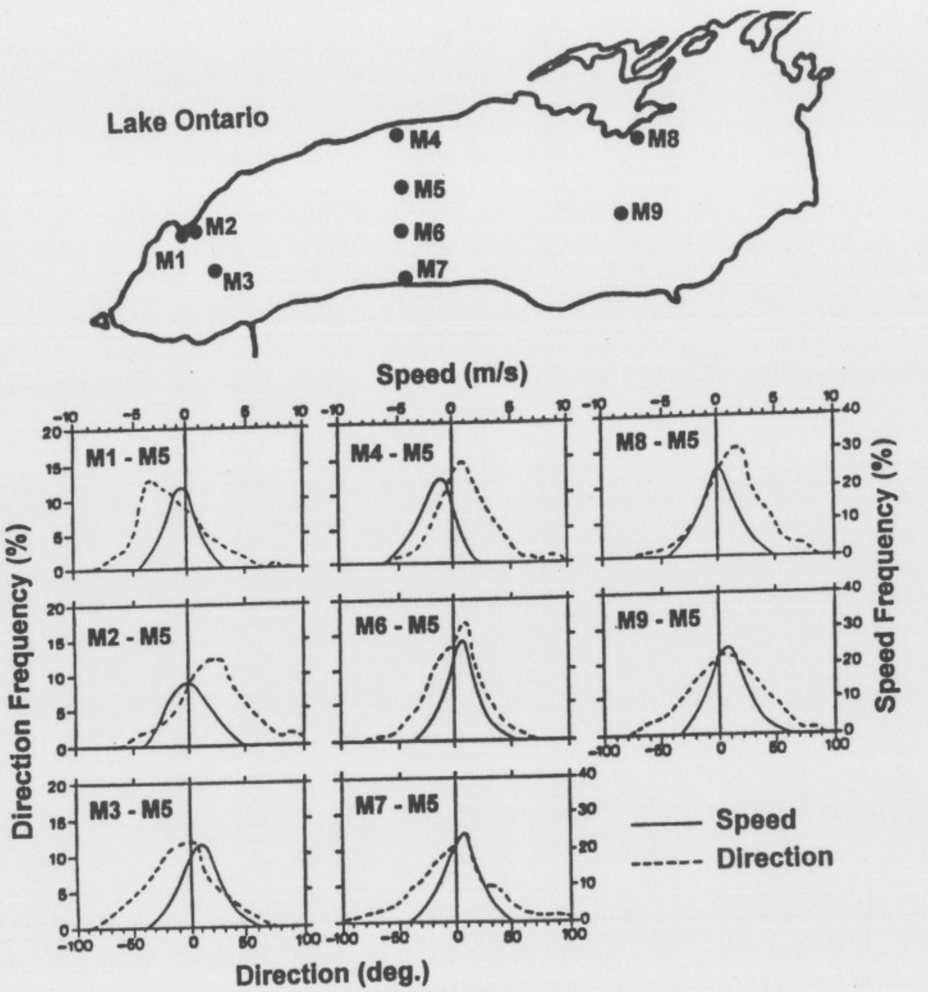

Figure 2-12. *Frequency distribution of wind deviations between selected meteorological stations on Lake Ontario for 6 May - 30 August 1982 (Simons and Schertzer, 1985)*

2.4.6 Humidity, vapor pressure and dew point temperature

Phillips and McCulloch (1972) examined the vapor pressure difference across the Great Lakes basin as a conservative measure of humidity and found pronounced seasonal, latitudinal, and longitudinal variations (Figures 2-14a to 2-14d). During winter (Figure 2-14a) vapor pressure isolines present a rather uniform gradient. water tends to increase the moisture content of downwind locations by 20 % over the upwind location values. The effect of ice cover (e.g., Lake Erie) is to minimize the lake as a source of moisture.

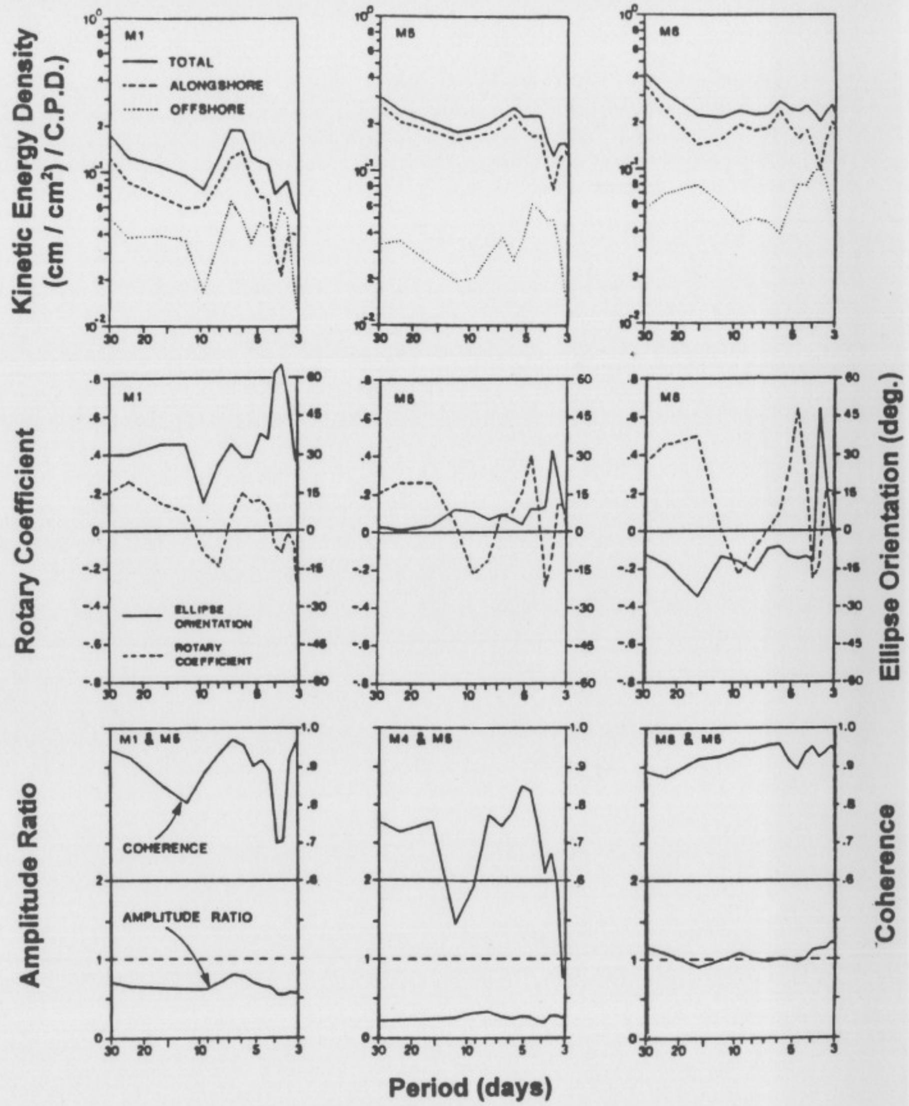

Figure 2-13. *Energy spectra of wind stress, rotary coefficient and ellipse orientation for selected stations and coherence and amplitude ratio between selected stations on Lake Ontario for 6 May - 30 August 1982 (Simons and Schertzer, 1985).*

During spring (Figure 2-14b), pressures were seen to vary between 4.5-8.5 mb with a remarkably uniform latitudinal gradient. Phillips and McCulloch (1972) indicated that the small contrast between spring air and water temperatures inhibits evaporation and downwind moisture transfer. Isolines during summer months are more complicated than during winter and spring varying by as much as 6 mb across the Great Lakes basin

(Figure 2-14c). During the fall months (Figure 2-14d) the lakes become an important source of moisture resulting in an increase of vapor pressure at downwind stations by 5-15 %. Diurnal variations of vapor pressure across the Great Lakes are generally small (i.e. < 0.8 mb).

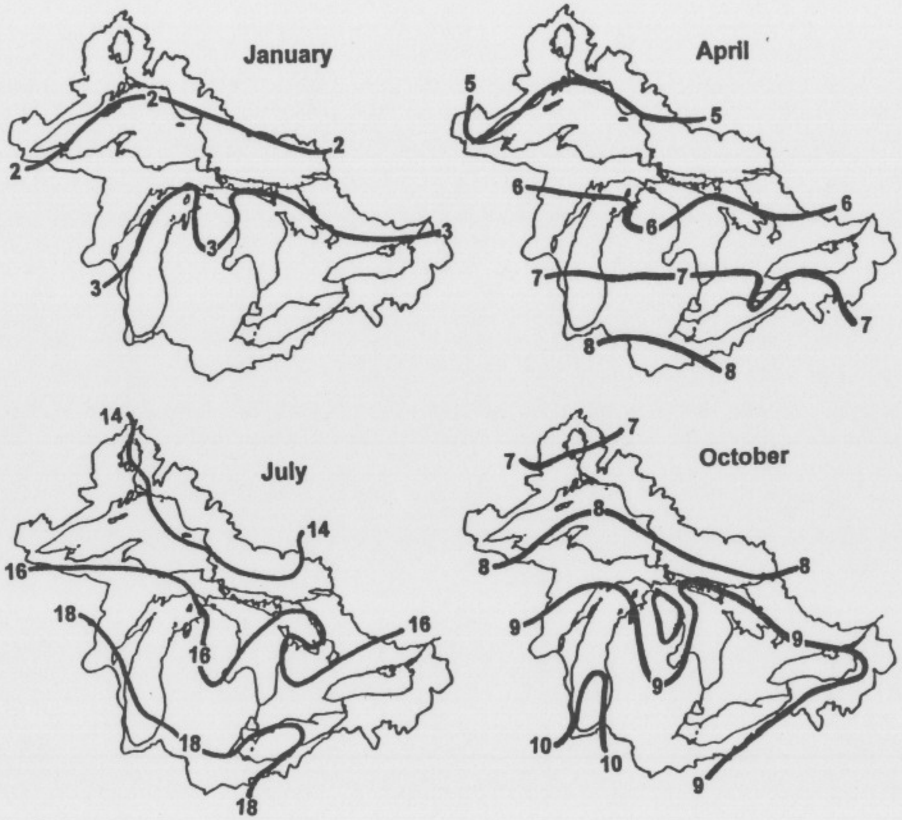

Figure 2-14. *Spatial variation of vapor pressure (mb) over the Great Lakes Basin (a) January, (b) April, (c), July and (d) October. (Phillips and McCulloch, 1972).*

Figure 2-15b illustrates mean, minimum, and maximum 5-day means of dew point temperature for Lake Erie over the period 1953-1983. Dew point temperature follows a similar seasonal variation as air temperature (Figure 2-15a). Minimum values are found in the winter months and peak values occur during the summer. The difference between maximum and minimum dew point observations is largest during the fall and winter months. As with other standard meteorological observations, dew point temperature and relative humidity observations are more spatially dense in the lower Great Lakes. Such observations have been important for long-term evaluation of lake evaporation.

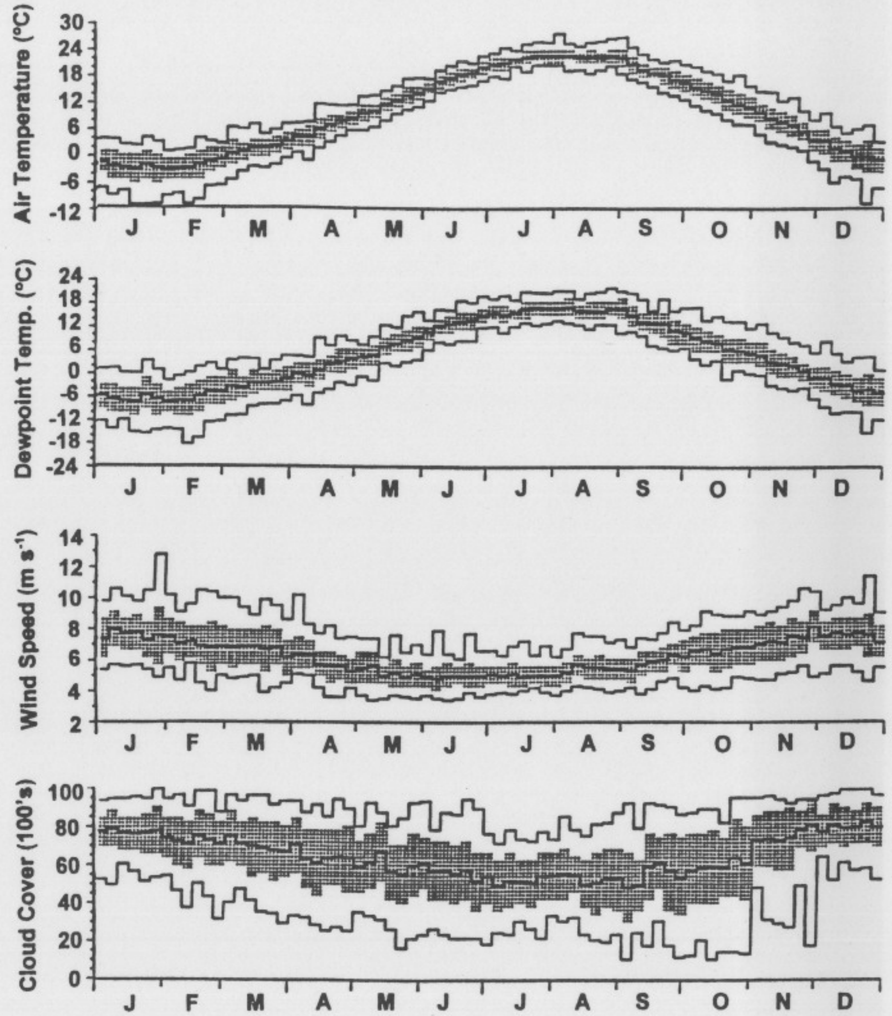

Figure 2-15. *Mean, maximum and minimum seasonal variation of (a) air temperature, (b) dew point temperature, (c) wind speed and (d) cloudiness for Lake Erie 1953-1983. (Schertzer et al., 1993)*

2.4.7 Cloudiness and fog

Cloud amount affects the receipt of global solar radiation at the earth's surface and as such has an effect on the basin hydrological balance as well as the lake thermal regime and water quality conditions. Figure 2-15d provides an example of the long-term mean, minimum, and maximum for 5-day mean cloudiness over Lake Erie. Cloudiness appears to have a seasonal component in that minimum values occur during the summer and early fall months and maximum values (mean > 75%) occur during the winter.

Advection of air across water can result in the formation of fog which can have a significant effect on radiation and energy budgets of lakes especially in the spring heating period. An analysis of long-term fog records (i.e. Phillips and McCulloch, 1972) for meteorological stations at the periphery of the lake showed that fog occurrence ranges from approximately 10 to 60 days per year. During IFYGL, it was estimated that the presence of fog over the lake modified the net radiation balance by as much as 30% highlighting the need for accurate determination of fog extent and density (Pinsak and Rodgers, 1981; Schertzer, 1982). Cold water and warm water advection fog are common on the Great Lakes. Under conditions of cold water advection fog, warm air approaching the shoreline from the offshore waters encounters the leading edge of near-shore cold up-welled water. Fog occurs if the air is cooled below the dew point. Under conditions of warm-water advection fog, cold air traverses a warmer water surface and the warmer, moister surface air is unstable and convects moisture into the cooler air. Condensation results in a fog pattern that looks like rising steam (Oke, 1983).

2.4.8 Ice cover

Extensive ice cover develops on most of the lakes during most winters. Figure 2-16 illustrates the long-term mean and range of ice extent for the Great Lakes (Assel et al., 1983). Lake Superior averages about 75% ice-covered, Michigan is 45%, Huron is 68%, Erie is 45%, and Ontario is 24%. The Great Lakes do not ordinarily freeze-over completely (Assel et al., 1983) because of their large heat storage capacity, large surface area, and their location in the mid-latitude winter storm track. Ice formation generally begins in the shallow shore areas of the Great Lakes in December and January. The

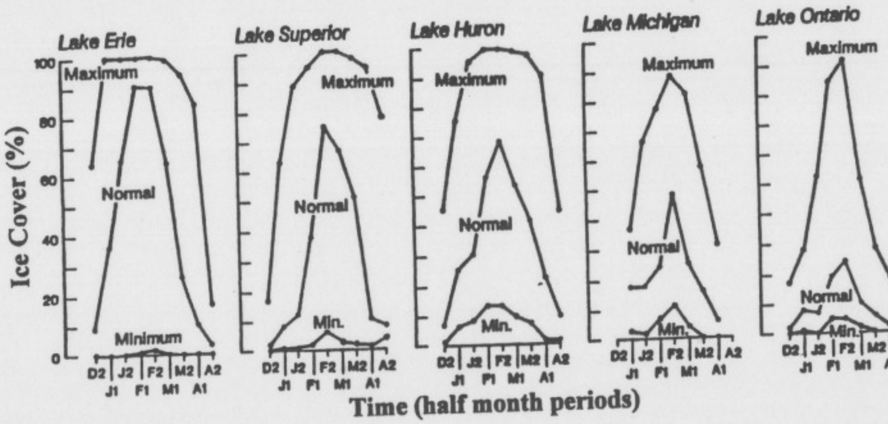

Figure 2-16. *Long-term maximum, minimum and normal ice concentration distribution patterns on the Great Lakes (Assel et al., 1983)*

deeper mid-lake areas normally do not form extensive ice cover until February and March. Ice is lost over all lake areas during the last half of March and during April. See Chapter 6 for detailed discussion of ice dynamics on the Great Lakes.

Transitory ice conditions result from air mass changes and episodic wind stress conditions. Ice cover in mid-lake regions is often in motion (Rondy, 1976). For example, Lake Erie ice speeds have been observed to average 8 cm s^{-1} with a maximum speed of 46 cm s^{-1} (Campbell et al., 1987). Ice movement can result in ice rafting forming rafted rubble 5-10 m thick.

2.5 REGIONAL CLIMATOLOGICAL MODELLING AND FLUCTUATIONS

The behavior of the Laurentian Great Lakes system is governed by its huge storages of water and energy. Climatological studies of the Great Lakes have concentrated on establishing the mass and energy balances on the basin and lakes. These dominant processes are largely responsible for the observed lake levels, storage changes, and the annual thermal cycle of the lakes.

The following section provides a brief background to regional climatological modeling concentrating on mass and energy balances of the Great Lakes system. Modeling results are dependent on the spatial and temporal adequacy of data networks and data bases as indicated above (Section 2.4). The Great Lakes basin has a wealth of meteorological, hydrological and limnological data; however, there are obvious deficiencies for long-term lake-wide (hydrodynamic or water quality) analyses. Some of the deficiencies may be alleviated through increased use of remotely sensed data (satellite) and also from application of mesoscale research results.

2.5.1 Mesoscale investigations

Broadly defined, climate includes the entire spectrum of weather in time and space. The integration over time of individual weather events characterizes a regional climate. Mesoscale (atmospheric phenomena on a scale larger than that of micrometeorology but smaller than the cyclonic scale) weather events are determined by the interaction of the large-scale flow with smaller-scale variations in topography and surface characteristics. With respect to large lakes such as the Great Lakes, land-water contrasts can be significant. Mesoscale atmospheric effects are important factors in understanding the climate of the Great Lakes.

Anthes (1992) suggests that it would be impossible to understand large-scale global climate without taking into account mesoscale atmospheric processes. Mesoscale weather includes such phenomena as hurricanes, severe thunderstorms, ice/snow storms, lake-land breezes, and so forth. Many of the exchanges of heat, moisture, momentum, and chemical trace species that determine global physical and chemical climates occur on the mesoscale. They occur in the surface and boundary layers of the atmosphere, in mesoscale frontal zones, and in cloud systems that range from individual cumulous clouds to mesoscale complexes of thunderstorms. Cloud-radiation interactions are amongst the most uncertain of the mesoscale physical processes which limit global climate models.

The IFYGL dense measurement network provided an opportunity to analyze some mesoscale lake effect processes over Lake Ontario both in time and space. Land station data as well as additional lake observations from buoys and towers were used in this study. Phillips and Almazan (1981) reported that the diagnostic analysis of three-dimensional motion, temperature, and moisture structure over the lake revealed a number of complex frontal structures that were not evident from synoptic-scale data alone. These included lake and land breeze circulation, lake effect snowstorms, and major storm and cold frontal passages. Such processes have a pronounced effect on local diurnal weather conditions, especially at the lake periphery. Major storms or frontal processes have an effect on basin hydrology, lake energy exchange, and thermal characteristics.

2.5.1.1 Lake and land breeze

Land and water surfaces possess contrasting thermal responses due to their different properties and energy balances. The difference in thermal environments is the driving force that results in the development of lake breeze circulation near the shoreline. Over a large water body the reduced convective heat flux to and from the air means that atmospheric warming and cooling rates are relatively small over water bodies (Oke, 1983). In contrast, the convective fluxes and rates of temperature change over land are large and show marked diurnal variation. These land-water temperature differences and their diurnal reversal (by day - land warmer than water; by night - land cooler than water) produce corresponding land water pressure differences which result in a system of breezes across the shoreline which reverse their direction between day and night. The lake and land breeze process also affects the temperature and humidity in the localized area on either side of the shoreline.

2.5.1.2 Lake effect snow storms

Numerical mesoscale models of lake effect have been developed to model lake effect snowstorms. Such storms commonly occur on the lee side of large lakes. Factors which affect the development of such storms include orographic lifting, surface friction, surface heating, evaporation and wind shear (Phillips and Almazan, 1981). Mesoscale analysis has indicated that surface heating plays a dominant role in causing convergence to occur which intensifies with the release of latent heat above the lake. In the case of Lake Ontario, the surrounding orography and vertical wind shear strengthens the disturbance causing precipitation to be distributed farther from the lake. Wind aligned along the axis of Lake Ontario results in a large fetch resulting in narrow, intense storms over the central part of the lake and the lee shore. Less intense lake effect storms occur when the wind fetch is shorter.

2.5.1.3 Major storm events

Occasionally large scale weather events such as hurricanes can impact on the Great Lakes region. One such event, Hurricane Agnes (20-25 June 1972) occurred during the IFYGL measurement program. Gale force winds were associated with the hurricane. Observed strongest winds were from the north and north east over Lake Ontario with speeds up to 22 m s^{-1}. Precipitation during the event ranged from 356 mm over the southwestern counties of New York state to less than 50 mm in the Toronto area. Maximum rainfall intensities exceeded all previous intensities for a duration of 3 hours or more for some stations at the southwestern part of New York state. For the 5-day storm, total rainfall ranged from 38-89 mm on the Ontario side of the basin while most of the land area south of the lake had rainfall more than 100 mm with amounts in excess of 250 mm being recorded in some areas. Storms of the intensity of Hurricane Agnes have the potential to provide very high discharge rates from tributaries to the lakes. In this storm, an instantaneous peak discharge of 984 m^3 s^{-1} was almost five times the mean annual maximum discharge of the Genesee River with a probability recurrence interval of 200 years. Effects on Lake Ontario included the generation of waves in excess of 4 m which resulted in shoreline erosion damage. Water temperatures were reduced over the storm event. Since the hurricane arrived during the critical spring heating period, existing unstable thermal stratification was disrupted by high wind mixing through the water column.

The circulation and mixing in the upper 10m of large lakes systems (i.e. Lake Ontario) is generally poorly known especially during major storm events. Hamblin et al. (1996) conducted time-series measurements of winds, air temperature, water temperature and current profiles from an inverted acoustic doppler profiler and fixed point moorings at an open lake location in Lake Ontario, which were instrumental in providing insight on the hydrodynamical responses during Hurricane Opal - one of the most severe storms in recent history in the area. Northeast winds started at the beginning of October 5, 1995 reaching a peak of nearly 18 m s^{-1} 24hr later slowly subsiding in the subsequent 12hr period. During the storm, currents continued as inertial or Poincare waves but at higher level reached a maximum of 50 cm s^{-1} seven hours after the peak wind speed with little intensification near the surface. Currents following the peak exhibited pronounced vertical shear with a jet developing at the thermocline 30 hours after peak winds. A train of large amplitude (10m) internal waves disturbed the thermal structure following the hurricane and persisted for several days.

2.5.1.4 Cold front passage

Advection of cold Arctic air over the Great Lakes can have a pronounced effect on meteorological conditions and on lake thermal structure. McBean (1975) described a cold front passage which occurred during IFYGL on October 8 to 10, 1972 with a frontal speed of 16 m s^{-1}. The surface front was complex and advanced toward Lake Ontario which had air and water temperatures between 15-16°C. The cold front resulted in a large change in atmospheric moisture content over Lake Ontario and precipitation. Lake surface temperature dropped from an average of 15.3°C to 11.5°C due to a direct loss of heat to the atmosphere and to mixing of cold hypolimnion water into the warm epilimnion. McBean (1975) had calculated that more than 60 % of the decrease in lake temperature between late September and late November was attributable to frontal passages with strong northwesterly winds which occupied less than 15 % of the time.

2.5.2 Lake heat budget and thermal responses

The thermal cycle and structure of the Great Lakes as well as biochemical components within the lake are influenced by meteorological, hydrological and limnological factors. Much research has been conducted to determine lake heat budgets which are important in finding lake heat storage, stratification characteristics, and vertical temperature structure.

2.5.2.1 Lake heat budget

The change in lake heat storage (Q_S; J m^{-2}) can be determined by using detailed observations of the vertical temperature structure between two survey periods,

$$dQ_S = V(T_2 - T_1)\,\frac{\rho\, cp}{A\,\Delta t} \tag{2-1}$$

where V = volume (m^{-3}), T = temperature (°C), A = surface area (m^2), ρ = density of water (mg m^{-3}), C_p = specific heat of water (J mg^{-1} °C), and Δt = time interval (Schertzer, 1997). The limiting factor is availability of detailed temperature profile measurements at adequate temporal and spatial resolution. The alternative is to observe or compute the other components of the lake energy budget and to determine the heat flux as a residual.

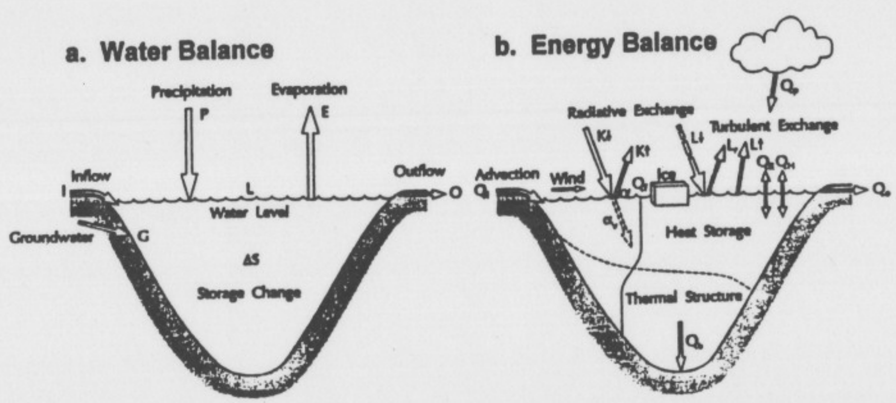

Figure 2-17. *Main components of the (a) lake hydrological and (b) lake energy balance.*

The main components of the energy balance of lakes is illustrated in Fig. 2-17a. The surface heat flux (MJ m^{-2} d^{-1}) can be expressed as the following basic heating or cooling processes,

$$Q_s = K\downarrow - K\uparrow + L\downarrow - L\uparrow - L_r - Q_e - Q_h - Q_I - Q_v \tag{2-2}$$

where $K\downarrow$ = incoming global solar radiation, $K\uparrow$ = reflected global solar radiation, $L\downarrow$ = incoming long-wave radiation, $L\uparrow$ = emitted long-wave radiation, L_r = reflected long-wave radiation, Q_e = latent heat flux, Q_h = sensible heat flux, Q_I = heat flux due to ice formation and decay, and Q_v = advected heat flux.

Because of the depth of the Great Lakes, other minor fluxes such as heat flux through the lake bottom are generally neglected (Rodgers, 1969). Formulations for these energy budget components are variously defined (e.g. Derecki, 1975; Schertzer, 1978, 1987; Schertzer and Sawchuk, 1990; Henderson-Sellers and Davies, 1989; Henderson-Sellers, 1986, 1990; Croley, 1989b). See the next section for an example Great Lakes thermodynamics model (Croley, 1989a,b, 1992a; Croley and Assel, 1994) especially suited for broad applications (long simulations of large areas) useful in climate change impact assessments.

Figure 2-18 illustrates long-term monthly mean surface heat flux determined for the Great Lakes by several researchers (Schertzer, 1997). Positive changes in lake heat storage (lake heat gains) generally occur from February to September primarily through net radiation heating as the turbulent exchange components are generally small at this time of year. During the lake cooling period in the fall, negative changes in the lake heat storage for the Great Lakes occur primarily through the turbulent exchange components since the net radiative flux is small during these months. Maximum heat losses occur December through January and maximum heat storage occurs in late August and early September coinciding with the period in which the surface heat flux turns negative.

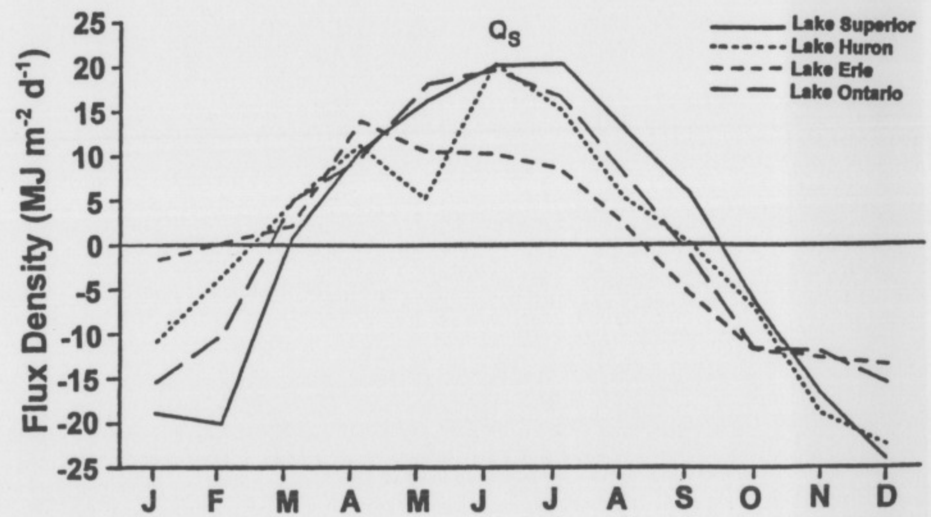

Figure 2-18. *Long-term monthly mean surface heat flux for the Great Lakes (Schertzer, 1997).*

Ice formation alters the surface thermodynamics of the lakes, changing subsequent ice formation, surface heating or cooling, lake evaporation, and lake responses to atmospheric changes. The large heat storages of the lakes provide a buffering; they forestall and reduce ice formation and shift the large evaporation response. Water temperatures lag air temperatures and evaporation lags surface heating (insolation). Evaporation peaks in October-November on Lake Erie and in November-December on Lake Superior.

Because of the high specific heat of water, large lakes store great quantities of heat and react slowly to short term changes in temperature. The heat storage of a large lake represents the integral of heating and cooling processes as a result of air/water interactions and hydrological balances. The annual heat storage can be divided into a summer heat income (the heat required to raise the lake temperature from 4°C to maximum temperature) and the winter heat income (the heat required to raise the lake temperature from minimum heat content up to 4°C). Long-term heat content varies widely among different lakes with good positive correlation with morphological characteristics such as lake area, depth and volume (Gorham and Boyce, 1989).

The largest annual heat budget for world lakes is that of Lake Baikal, Russia (approximately 3.1 GJm^{-2}) (Hutchinson, 1957). Figure 2-19 shows a comparison of heat incomes for the Great Lakes. Analysis of the thermal regime of Lake Superior (Bennett, 1978; Schertzer, 1978) indicated that the average total Spring heat income of 1.47 GJm^{-2} accounted for a rise in mean lake temperature from a minimum of 1.4°C to the average temperature of maximum density 3.83°C (temperature of maximum density is affected by pressure) and decreases 0.1°C for each 100 m depth) while the average total Summer income of 1.26 GJ m^{-2} accounts for the remaining rise to the maximum mean lake temperature of 5.88°C. For Lake Superior, approximately 54% of the annual heat income is used for spring warming of the water to maximum density. In general,

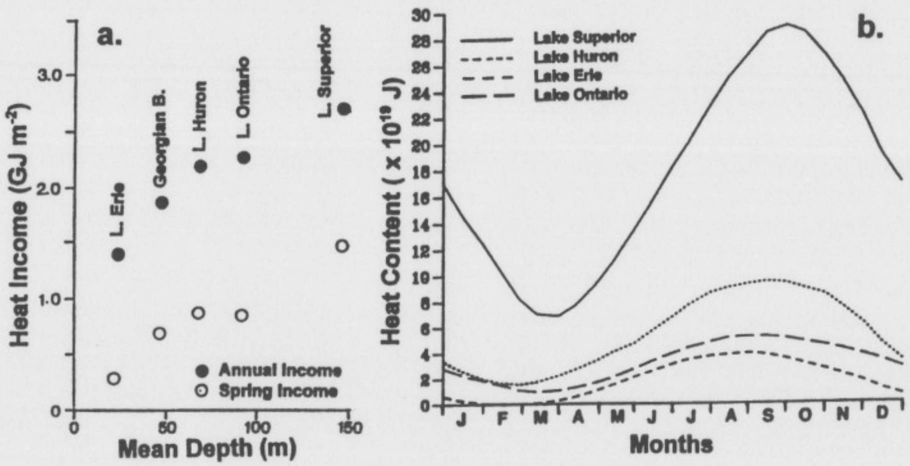

Figure 2-19. *Comparison of (a) heat incomes for the Great Lakes (Bennett, 1978) and, (b) heat content of the Great Lakes (Schertzer, 1997).*

Figure 2-19 shows that as lake surface area, depth, or volume increases, there is a rise in heat uptake and storage. The large heat incomes for the Great Lakes is related to the "dimictic" nature of these lakes, which allows the entire lake volume to be involved in heat exchange semi-annually. The seasonal cycle of heat content for the Great Lakes is shown in Figure 2-19b, estimated from Boyce et al. (1976); Schertzer (1978, 1987); and Bolsenga (1975). On the Great Lakes, minimum heat storage occurs in late winter while maximums occur in late summer/early fall. Substantial differences in the timing of maximum and minimum heat content in each lake is related to the variation in volume.

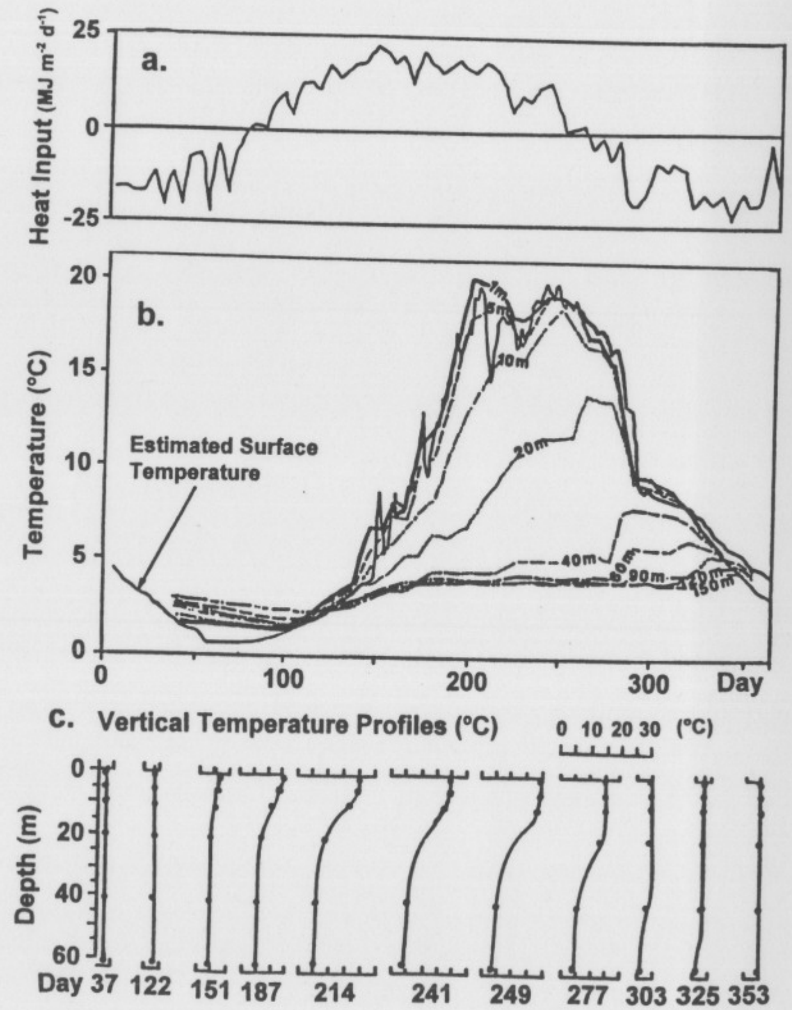

Figure 2-20. *Thermal response of Lake Ontario during IFYGL (a) heat input, (b) temperature at selected depths and (c) lake-wide mean vertical temperature profiles. (Based on IFYGL investigations - IFYGL, 1981).*

2.5.2.2 climatic effect on large lake annual thermal cycle

The Great Lakes are subject to major seasonal changes in the net heat input and as a consequence go through an annual thermal cycle. Under current climatic conditions, the Great Lakes are dimictic (mix from top to bottom twice yearly in the spring and the fall). The timing of the overturn is very closely related to the time when the surface water temperatures of the lake fluctuates through the temperature of maximum density of fresh water, (i.e. 4°C).

In the winter, the surface waters are all generally below the temperature of maximum density. In the spring, progressive warming of the near-shore waters reach temperatures above 4°C while waters immediately offshore are still below 4°C and a zone of convergence referred to as the "thermal bar" is initially formed along the near-shore marking the onset of summer stratification. Progressive heating advances the thermal bar toward the center of the lake, and when the mid-lake reaches 4°C, the thermal bar disappears. In deep lakes such as Lake Ontario, Huron and Michigan, this process takes as much as six to eight weeks. Typical summer stratification is achieved as the surface waters reach greater than 4°C over the entire lake.

Although thermal stratification proceeds at varying rates over the Great Lakes, the rate of heat input is generally maximum in the mid-summer period (Figure 2-18) and thermal stratification has become generally established over the entire lake. Figures 2-20a to 2-20c illustrate calculated heat input and measured temperature response of Lake Ontario during IFYGL. In late spring, a thermocline is first formed close to the surface and gradually deepens as the heat gains continue to exceed heat losses in surface waters. Lake averaged vertical temperature profiles for Lake Ontario (Figure 2-20c) illustrates progressive deepening of the mean thermocline level as well as the formation of the warm upper well mixed epilimnion layer, the deep cold hypolimnion, and the transition zone referred to as the mesolimnion layer.

Toward the end of the summer, the lakes have attained maximum heat content and by early fall the mean heat content starts declining due to autumn cooling. By mid-fall, the autumn cooling and associated mixing processes have nearly completed the breakdown of thermal stratification. With further cooling, the depth of the mixing deepens until the entire water column is mixed around 4 to 5°C. Although horizontal temperature gradients persist close to the shore, vertical mixing of the open water is nearly completed by late fall. This is commonly referred to as the annual fall overturn. With continued cooling coupled with wind mixing, the main water mass continues to be well mixed during the winter, attaining isothermal conditions at the temperature of maximum density, 4°C, by mid-winter. Higher cooling rates near-shore result in horizontal temperature gradients which persist throughout the winter. During late fall and early winter, mixing of cold inshore water with warmer offshore water may set up a convective regime (thermal bar) described earlier. Towards the end of winter, the entire water mass has cooled down to below 4°C, with the coldest water remaining close to the shore.

2.5.2.3 Climate (heating and wind) effect on thermal stratification

The seasonal cycle of thermal development for a mid-lake station in Lake Erie during the spring, summer and fall months of 1979 was described by Schertzer et al. (1987) in response to surface heating and wind mixing. During May a developing weak unstable stratification can be disrupted by storms. A stable thermocline is expected to develop during June. Major wind events can result in upper layer mixing and thermocline deepening. An example of thermal characteristics is shown in Figure 2-21 for July through September.

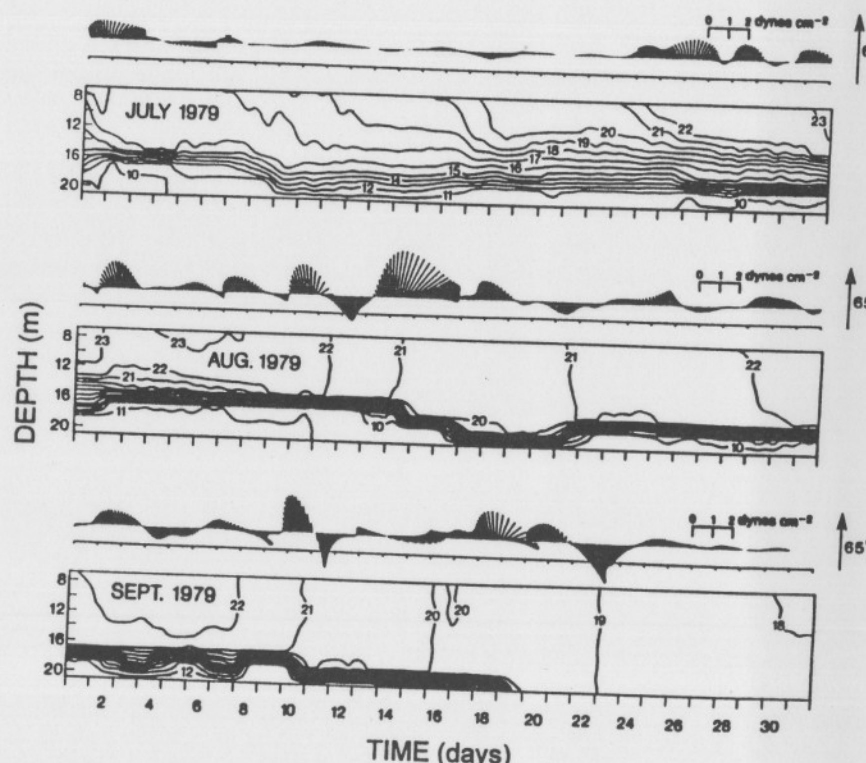

Figure 2-21. *Isotherms (°C) drawn from thermistor chain recordings at a mooring in the central basin of Lake Erie, May to September 1979 with low-pass filtered wind stress vectors (Schertzer et al., 1987).*

During July with continued heating and few major wind events, the stratification intensifies as gradients across the thermocline region increase. Hypolimnetic entrainment was seen to occur at the beginning of July (Ivey and Boyce, 1982). August 1979 was a month of intensifying thermocline gradients coupled with upper layer deepening and clear definition of a hypolimnetic layer 3-6 m thick. Major wind stress impulses during the month were observed to contribute to temperature gradients as high as 6°Cm^{-1}. September isotherms marked the end of the 1979 central basin stratification as the hypolimnion was eradicated on September 11, 1979 in response to wind-driven mixing associated with high wind stress impulses. Additional discussion can be found in Section 3.1 and Section 7.3.2.

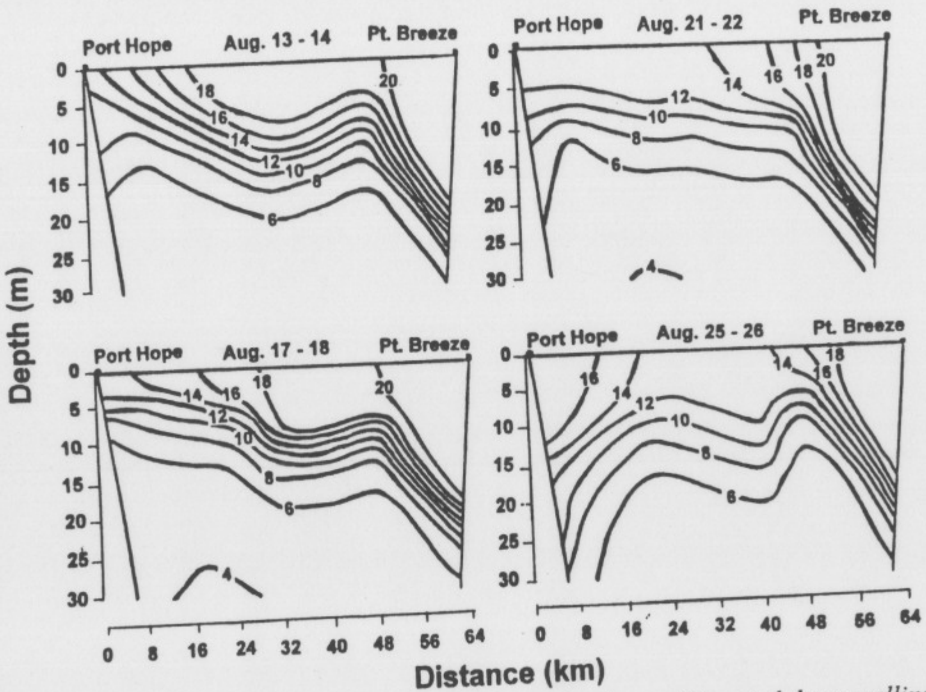

Figure 2-22. *Isotherms (°C) illustrating an episode of up-welling and down-welling in Lake Ontario in response to wind forcing along a transect from the north shore (Port Hope) to the south shore (Point Breeze). (Simons and Schertzer, 1987)*

2.5.2.4 Climate (wind) forced up-welling and down-welling events

Simons and Schertzer (1987) observed that at opposing shores of large lakes such as Lake Ontario, the thermocline moves in opposite directions clearly forced by the wind. An eastward wind causes southward surface drift and hence up-welling at the north shore and down-welling at the south shore; a westward wind has the opposite effect. From a water quality perspective, up-welling of hypolimnetic water at the shoreline is a mechanism for redistribution of nutrients in the water column. Changes in wind speed and direction (under climate change) may have the effect of altering the frequency of occurrence of up-welling events.

Figure 2-22 shows vertical temperature isotherms along a transect from the north shore (Port Hope) to south shore (Point Breeze) of Lake Ontario (Simons and Schertzer, 1987). Clearly depicted is a sequence of the development and decay of an episode of up-welling along the north shore of the cross-section with corresponding down-welling of warm water along the south shore. Additional discussion on lake hydrodynamics is provided in Chapter 4.

2.5.3 Lake levels and flows

Water quantity considerations have been a dominant area of investigation within the Great Lakes Basin and the lakes. Water quantity impacts on all sectors of the economy including concern for ecosystems (wetlands), navigation, hydroelectric power generation as well as recreation.

2.5.3.1 Great Lakes water balance

The main components of the hydrological balance of a large lake system are illustrated in Figure 2-17b. A generalized water budget for a lake system can be formulated over any time period in terms of equivalent water depths over the lake surface as follows:

$$dW = Q_I + P + R - E - Q_O \qquad (2\text{-}3)$$

where dW = water level change, Q_I = inflow, P = over-lake precipitation, R = runoff to the lake, E = lake evaporation, and Q_O = outflow. Table 2-3 provides a summary of the long-term water balance computed for the Great Lakes over 1951-1988.

Lake	Precipitation on the lake	Runoff to the lake	Evaporation from the lake
Superior	82	62	56
Michigan	83	64	65
Huron	87	84	63
Erie	81	80	90
Ontario	93	169	67

Table 2-3. *Great Lakes water balance, 1951-1988 (cm) (Croley 1995)*

2.5.3.2 Lake level fluctuation

Long-term annual lake levels, seasonal lake levels, and short-period lake level changes due to wind setup and storm surge are the primary lake level fluctuations. Figure 2-23 illustrates long-term water levels for the Great Lakes.

Annual lake level fluctuations are primarily responsible for much of the variability in water levels (record highs in 1952, 1973 and 1986 and record lows in 1935 and 1964). From 1960 to the present there is an overall range of about 2 m in the annual levels. Continuing high water levels in the Great Lakes are the result of the increased precipitation regime since 1940 coupled with the lower temperature regime since 1960 (Croley, 1995).

Seasonal water level changes range from about 30 cm on the upper lakes to about 38 cm on the lower lakes. The cycle begins with minimum water levels generally during winter. Spring snow melt and precipitation contribute to raising levels to seasonal maximums which occur in June for the smaller lakes, Erie and Ontario, and September in the case of Lake Superior. Decreasing net water supplies in summer and fall result in a progressive seasonal decline to winter minimums.

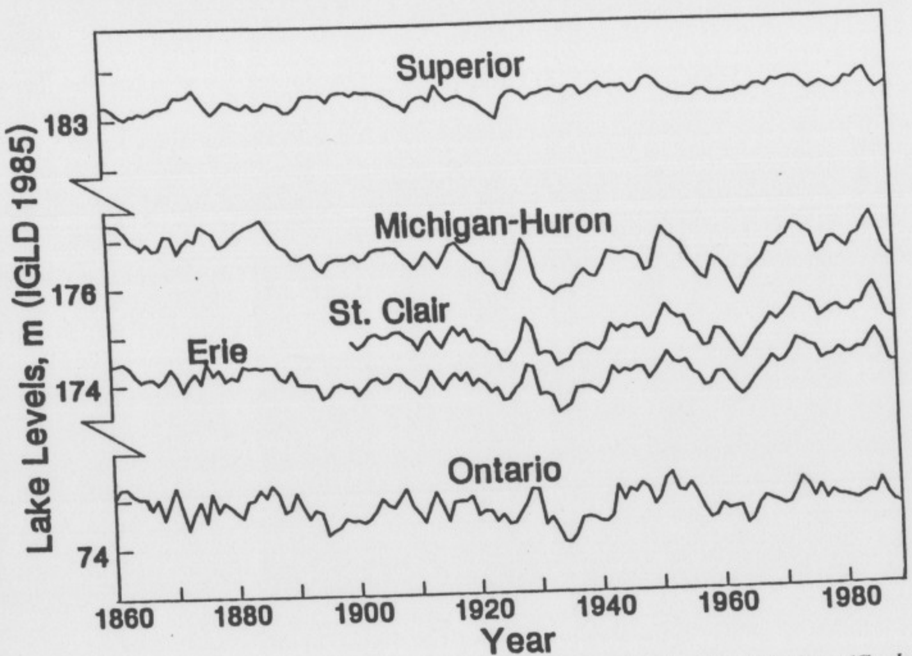

Figure 2-23. *Long-term annual water level fluctuations for the Great Lakes. (Croley, 1995).*

Storm surges and wind setup are common fluctuations especially in shallower lakes and embayments (i.e. Lake Erie, Saginaw Bay) when strong persistent winds blow along the long axis of such lakes. Rapid differences in levels between one end of the lake and the other can be as extreme as 5 m (Lake Erie 2 December 1985) and can be responsible for most of the Great Lakes shoreline damage (Hamblin, 1979).

2.5.3.3 Connecting channel flows, diversions and flushing time

Physical aspects of connecting channel flows and diversions in the Great Lakes system are discussed in Section 2.2 and Table 2-2a provides a summary of the long-term mean and range of flows. As indicated in Table 2-2a, average connecting channel discharges progressively increase downstream from 2100 $m^3 s^{-1}$ at the St. Marys River to 6700 $m^3 s^{-1}$ at the St. Lawrence River.

With respect to the effect of diversions on the Great Lakes, the Long Lac and Ogoki Diversions average about 160 $m^3 s^{-1}$ and raise lake levels between 6 cm and 11 cm. The Chicago Diversion averages about 90 $m^3 s^{-1}$ and lowers lake levels between 2 cm and 6 cm. The Welland Canal, which bypasses Niagara Falls, averages about 270 $m^3 s^{-1}$ and lowers lake levels between 2 cm and 13 cm with no effect on Lake Ontario. The combined effect on the lakes ranges from a 2 cm rise for Lake Superior to a 10 cm drop for Lake Erie. The diversion effects are therefore small in comparison with the one or more meter variation associated with short-term storm movements, the 30-38 cm seasonal cycle, and the 2 m range of annual variations. Present-day diversions have a relatively small effect on the Great Lakes. Combined with the long response time of the system, diversions are not suitable for lake regulation.

Mean annual outflows and lake volumes can be combined to yield a water flushing time. The flushing time, which is of interest for water quality studies and predictions of the average length of time that molecules or ions of a dissolved, non-reactive constituent of the lake water remain in the lake before being transported out by the flowing rivers; the flushing time, therefore, is the lake volume divided by the rate of outflow. Estimates of the flushing times in years for the Great Lakes are as follows, Lake Superior (165 yrs.), Lake Michigan (69.5 yrs.), main Lake Huron (10.6 yrs.), Georgian Bay (5.7 yrs.), Lake Erie (2.5 yrs.), Lake Ontario (7.5 yrs.). Although there are various estimates for the actual flushing times based on different estimates of volumes and discharge rates, the long flushing times for water in Lake Superior and Lake Michigan indicates that the recovery from an undesirable state of water quality might take a long time.

2.5.4 Great Lakes Hydrology and Modeling

As indicated above, measurement networks include basic observations of precipitation, and lake water level. Hydrological models have been developed and applied within the Great Lakes basin to estimate components of the hydrological cycle to provide information required for managing water supply problems of the region.

Great Lakes Environmental Research Laboratory (GLERL) applied hydrological process models on the Laurentian Great Lakes (including Georgian Bay and Lake St. Clair, both as separate entities) for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1993a,b; Croley and Hartmann, 1987, 1989; Croley and Lee, 1993; Hartmann, 1990). The hydrological models include rainfall-runoff [121 daily watershed models (Croley, 1982, 1983a,b; Croley and Hartmann, 1984)], over-lake precipitation (a daily estimation model), one-dimensional (depth) lake thermodynamics [7 daily models for lake surface flux, thermal structure, and heat storage (Croley, 1989a,b, 1992a; Croley and Assel, 1994)], channel routing (connecting channel flows, outlet works, and lake levels (Hartmann, 1987, 1988; Quinn, 1978), lake regulation (International Lake Superior Board of Control, 1981, 1982; International St. Lawrence River Board of Control, 1963), and diversions and consumption (International Great Lakes Diversions and Consumptive Uses Study Board, 1981).

2.5.4.1 Runoff modeling

One example of rainfall-runoff modeling on the Great Lakes is the GLERL Large Basin Runoff Model (LBRM), which consists of moisture storages arranged as a serial and parallel cascade of "tanks" (Croley, 1983a,b). Water flows from the snow pack to the upper soil zone tank, from the upper to the lower soil zone and surface storage tanks, from the lower to the groundwater and surface tanks, from the groundwater to the surface tank, and from the surface tank out of the watershed. It makes use of physical concepts for snow melt and net supply to the watershed surface, infiltration, heat available for evapotranspiration, actual evapotranspiration, and mass conservation. As a conceptual model, the LBRM is useful not only for predicting basin runoff, but to facilitate understanding of watershed response to natural forces as well. The main mathematical feature of the LBRM is that it may be described by strictly continuous equations; none of the complexities associated with inter-tank flow rate dependence on partial filling are introduced. For a sufficiently large watershed, these nuances are not observed due to the spatial integration of rainfall, snow melt, and evapotranspiration processes.

Daily precipitation, temperature, and insolation (the latter available from climatological summaries as a function of location) may be used to determine snow pack accumulations and net surface supply based on degree-day determinations of snow melt. The net surface supply is divided into infiltration to the upper soil zone and surface runoff by taking infiltration proportional to the net surface supply rate and to the areal extent of the unsaturated portion of the upper soil zone. Outflow from each storage within the watershed is proportional to the moisture in storage. The evapotranspiration rate from the upper and lower soil zones is proportional to available moistures there and to the heat rate available for evapotranspiration; it also reduces the heat available for subsequent evapotranspiration. The total amount of heat in a day is split between that used for, and that still available for, evapotranspiration by empirical functions of air temperature based on a long-term heat balance. Mass continuity yields a first-order linear differential equation for each of the moisture storages (Croley, 1982), which are tractable analytically; they are solved simultaneously to determine daily moisture storage, evapotranspiration, and basin runoff from daily data.

The Great Lakes basin is divided into 121 watersheds, each draining directly to a lake, grouped into the six lake basins. The meteorological data from about 1,800 stations about and in the watersheds are combined through Thiessen weighting to produce areally-averaged daily time series of precipitation and maximum and minimum air temperatures for each watershed (Croley and Hartmann, 1985). Records for all "most-downstream" flow stations are combined by aggregating and extrapolating for ungaged areas to estimate the daily runoff to the lake from each watershed. The LBRM is calibrated to determine the set of parameters resulting in the smallest sum-of-squared-errors between model and actual daily flow volumes for the calibration period (Croley, 1983b, Croley and Hartmann, 1984). After the LBRM is calibrated for each watershed, the model outflows are combined to represent each Great Lake basin; this distributed-parameter model integration filters individual sub-basin model errors. The LBRM calibration periods generally cover 1965-1982 depending upon flow data availability. The LBRM was also used in forecasts of Lake Superior water levels (Croley and Hartmann, 1987), and comparisons with climatic outlooks showed the runoff model was very close to actual runoff (monthly correlation of water supply were on the order of 0.99) for the period August 1982 - December 1984 which is outside of and wetter than the calibration period (Croley and Hartmann, 1986). The model also was used to simulate flows for the time period 1956-63, outside of the period of calibration. The correlation of monthly flow volumes between the model and observed values during this verification period are a little lower than the calibration correlation but quite good except for Lakes Superior and Huron (there were less than two-thirds as many flow gages available for 1956-63 as for the calibration period for these basins; the paucity of data making comparisons during this period more difficult compared to the other lakes).

2.5.4.2 Precipitation.

A lack of over-lake precipitation measurements means that estimates typically depend on land-based measurements and there may be differences between land and lake meteorology. Although gage exposures may significantly influence the results of lake-land precipitation studies (Bolsenga,, 1977, 1979), Wilson (1977) found that Lake Ontario precipitation estimates based on only near-shore stations averaged 5.6% more during the warm season and 2.1% less during the cold season than estimates based on stations situated in the lake. By using a network that also included stations somewhat removed from the Lake Ontario shoreline, Bolsenga and Hagman (1975) found that eliminating several gages not immediately in the vicinity of the shoreline increased over-lake precipitation estimates during the warm season and decreased them during the cold season. Thus, for the Great Lakes, where lake effects on near-shore meteorology are significant and the drainage basins have relatively low relief, the use here of all available meteorological stations throughout the basin is probably less biased than the use of only near-shore stations. Over-lake precipitation is taken equal to overland precipitation (on the basis of depth) without further corrections.

2.5.4.3 Lake thermodynamics

One example of a thermocline thermodynamics model for the Great Lakes is useful to consider here. It is a broad model, usable for long-periods (climate simulations) at a daily time interval. Chapter 3 contains a description of a model for detailed “short-period” simulations at a smaller-scale more suitable for lake hydrodynamic modeling. Broad models of the type described here use mass transfer formulations that include atmospheric stability effects on the bulk transfer coefficients, applied to monthly data for water surface temperatures, wind speed, humidity, and air temperatures (Quinn, 1979). As an example, GLERL uses that approach applied to daily data but combined with models for over-water meteorology, ice cover, and lake heat storage and with a lumped representation of a lake's heat balance (Croley, 1989a,b, 1992a). As over-water data are not generally available, GLERL uses over-land data by adjusting for over-water conditions. Phillips and Irbe's (1978) regressions for over-water corrections are used directly by replacing the fetch (and derived quantities) with averages. Air temperatures and specific humidities over ice are used for over-ice evaporation calculations and over water for the over-water calculations; the two estimates are combined by weighting for the fraction of the surface covered in ice. Water and ice pack heat balances (Croley and Assel, 1994) are used to relate ice cover extent to meteorology, heat storage, and surface fluxes between the atmosphere, the water body, and the ice pack.

Kraus and Turner's (1967) mixed-layer thermal structure concept is extended for the Great Lakes to allow the determination of a simple one-dimensional model for surface temperature increments or decrements from past heat additions or losses, respectively (Croley, 1989a,b, 1992a). The effects of past additions or losses are superimposed to determine the surface temperature on any day as a function of heat in storage; each past addition or loss is parameterized by its age. Turnovers (convective mixing of deep lower-density waters with surface waters as surface temperature passes through that at maximum density) can occur as a fundamental behavior of this superposition model, and hysteresis between heat in storage and surface temperature, observed during the heating and cooling cycles on the lakes, is preserved.

Heat in storage in the lake at the end of each day is given by a simple conservation of energy by taking the change in storage equal to the sum of the fluxes integrated over the day. As summarized by Gray et al. (1973), short-wave radiation is interpolated from generalized maps of Canadian and northern U.S. mid-monthly clear-sky values and adjusted for cloud cover. Average short-wave reflection is taken simply as one-tenth of the incident or as a function of ice cover, and sensible heat transfers at the water or ice temperature (minimum of air temperature or freezing temperature) are computed directly from the same mass transfer formulation and assumption (that the bulk evaporation coefficient is equal to the sensible heat coefficient) that is used to derive evaporation. It is then added to evaporative advection and latent heat transfers. Evaporative heat transfers from ice include the heat of fusion as well. Net long-wave radiation exchange is derived from considerations of the water and atmosphere as gray bodies with correction for cloud cover only to atmospheric radiation (Keijman, 1974). Net long-wave radiation exchange over ice is computed as for open water, ignoring the small effects of the ice surface on the exchange. Energy advected with precipitation is adjusted if the precipitation is snow, to account for the heat required in snow melt. Energy advected with precipitation onto the ice surface is uncorrected for melt since that is taken as occurring with ice melt, which is added to the budget when it happens. The energies advected into and out of the lake with other mass flows are relatively very small and are ignored. The equations representing evaporation, heat storage, and heat fluxes are solved simultaneously with daily data on over-land wind speed, air temperature, cloud cover, and humidity; details of an iterative solution technique are available elsewhere (Croley, 1989a,b, 1992a).

Unfortunately, there are no really good independent evaporation data to calibrate and verify evaporation models on the Great Lakes. Water balances are insufficient due to the large errors induced by subtracting nearly equal large inflows and outflows to each Great Lake, or due to errors in estimates of the water balance components. However, with the joint heat balance and evaporation model, it is possible to compare water surface temperatures with data, now available from the National Oceanic and Atmospheric Administration's Polar Orbiting Satellite Advanced Very High Resolution Radiometer (Irbe et al., 1982; AES, 1988).

Daily meteorological over-land data at from five to seven near-shore stations about each Great Lake were assembled and averaged for correction to over-lake data. The heat balance model was calibrated to give the smallest sum-of-squared-errors between model and actual daily water surface temperatures observed by satellite during the calibration period of generally 1979-88. There is good agreement between the actual and calibrated-model water surface temperatures; the root mean square error is between 1.1-1.6°C on the large lakes [within 1.1-1.9°C for an independent verification period, 1966-78 (Croley, 1989a,b, 1992a)].

2.5.4.4 Great Lakes net basin supply

The magnitude of the hydrological variables vary with season. An example with respect to Lake Erie is illustrated in Figure 2-24 (Quinn, 1982; Quinn and Kelley, 1983) based on evaluation of net basin supplies. In Lake Erie, the monthly precipitation is fairly uniformly distributed throughout the year, while the runoff has a peak during the spring which results primarily from the spring snow melt. The runoff is at a minimum in the late summer and early fall due to large evapotranspiration from the land basin. The lake evaporation reaches a minimum during the spring and gradually increases until

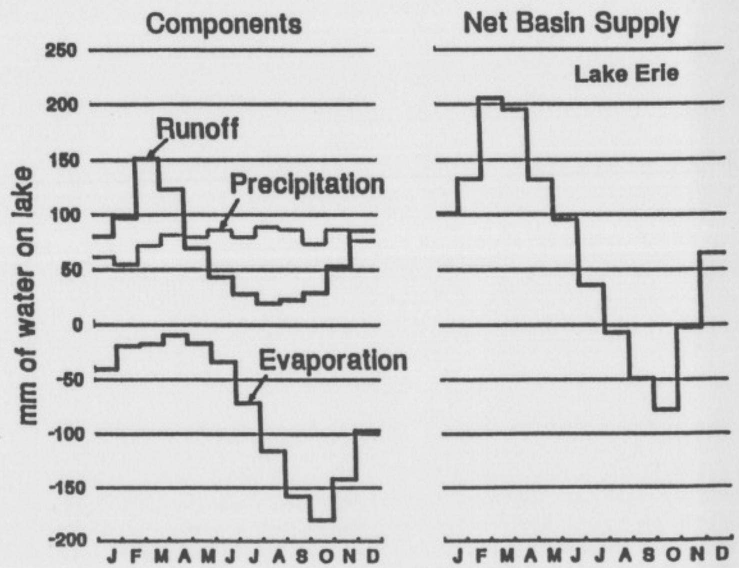

Figure 2-24. *Seasonal variation of precipitation, runoff and evaporation for Lake Erie based on evaluation of net basin supplies. (Quinn, 1982; Quinn and Kelley, 1983)*

it reaches a maximum in the late fall or early winter. The high evaporation period is due to very cold dry air passing over warm lake surfaces.

Management of the Great Lakes requires accurate estimates of the net basin supply (NBS). Lee (1993) indicates that NBS may also be an indicator of climatic change. The net basin supply refers to the net supply of water to the lake from its own basin. As outlined by Lee (1993), NBS can be computed either directly based on computer estimates of the hydrological components (i.e. Croley and Lee, 1993),

$$NBS = P + R - E \qquad (2\text{-}4)$$

where, NBS is net basin supply, P = over-lake precipitation, R = basin runoff and E is lake evaporation, or can be estimated as a residual from measured hydrological components,

$$NBS = \Delta S - I + O \pm D \qquad (2\text{-}5)$$

where, S = change in storage, I = inter-basin inflow through a natural channel, O = inter-basin outflow through a natural channel and D = inter-basin diversion into or out of the lake. Lee (1993) performed a statistical comparison between the direct and an

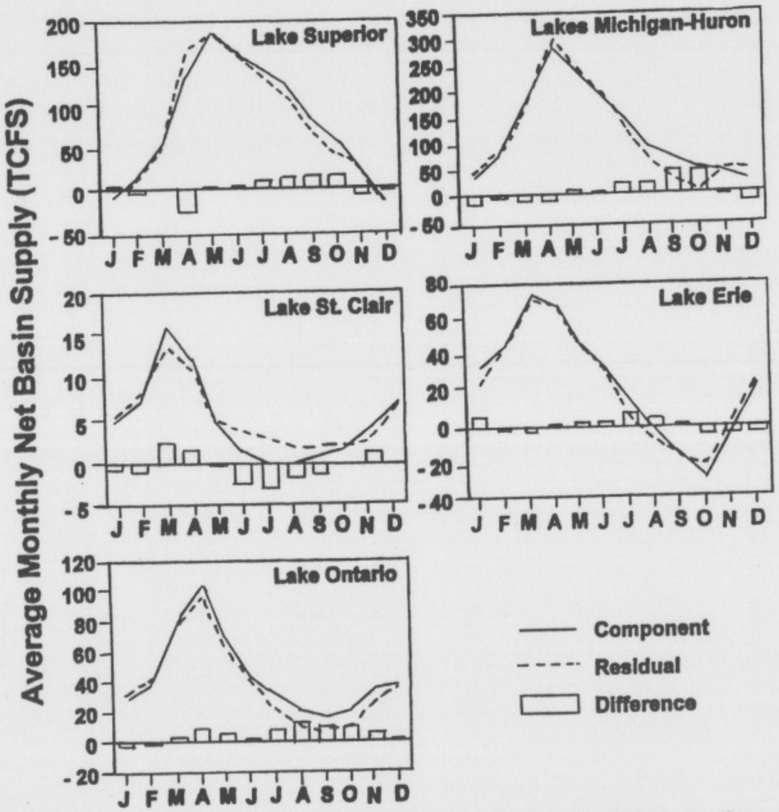

Figure 2-25. *Monthly average net basin supplies for the Great Lakes. (Lee, 1993)*

indirect approach to estimating NBS and found good agreement with high correlation (Lake Superior r = 0.96, Lake Michigan-Huron r = 0.97, Lake St. Clair r = 0.88, Lake Erie r = 0.98 and Lake Ontario r = 0.98). An evaluation of the differences indicated that for the residual approach, errors can be largely attributed to measurement errors. Relatively small measurement errors of large components can result in larger errors in the NBS. Errors in the direct computation of NBS components are related to errors in estimation of spatial values of precipitation, runoff and evaporation from few observations. It is estimated that for runoff, the percentage of the basin area gauged is 66% for Lake Superior, 76% for Lake Michigan, 57% for Lake Huron, 50% for Lake St. Clair, 78% for Lake Erie and 75% for Lake Ontario (Lee, 1993).

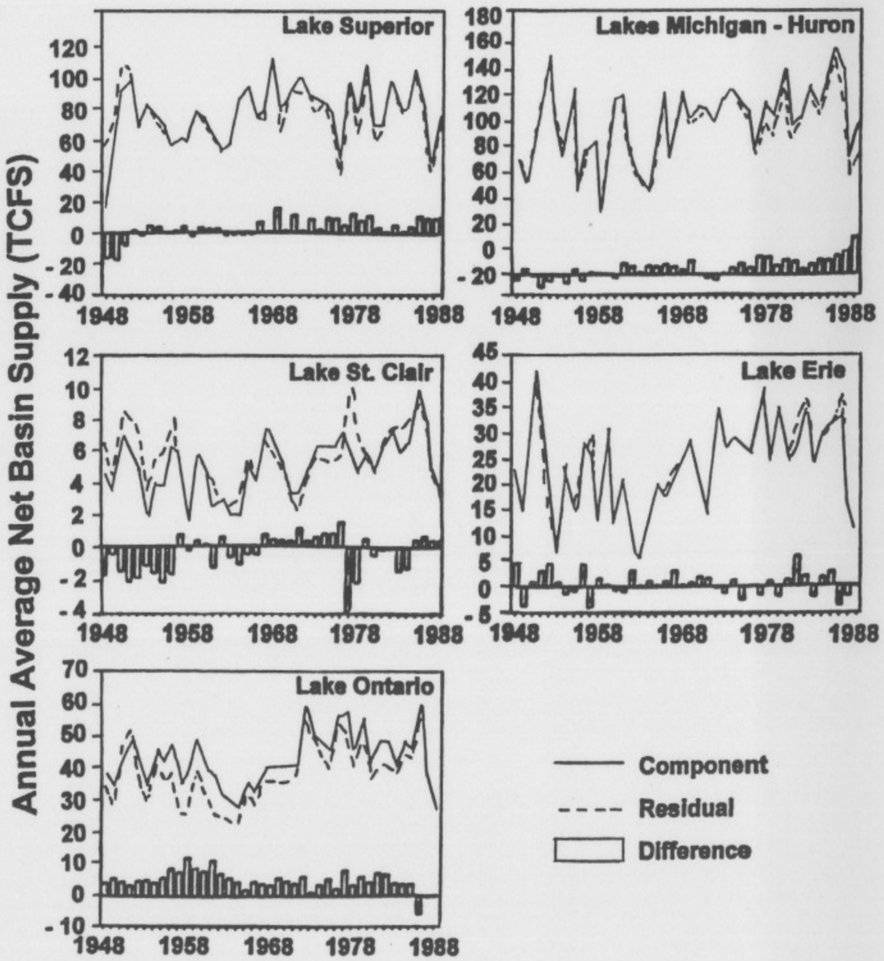

Figure 2-26. *Annual average net basin supplies for the Great Lakes (1948-1988). (Lee, 1993)*

As is apparent from Figure 2-25, the NBS varies considerably in magnitude for each of the Great Lakes. In general, the NBS is highest during the spring months and is minimum during the fall. Lake Superior, the upstream lake, exhibits a seasonal cycle which is less characteristic than that of the downstream lakes. Lee (1993) indicates that there is seasonal correspondence between the two methods of determining NBS, however, monthly means can be different and largest differences occur during the summer and fall months. Significant variability in the annual averages of the NBS is evident especially for Lakes Michigan-Huron and Lake Erie. NBS is largest for the Michigan-Huron lakes (Figure 2-26).

2.5.4.5 Lake levels and flows

Great Lakes levels and flows have been simulated for a variety of studies, including changed climates, (Quinn, 1988; International Joint Commission, 1976; Hartmann, 1990; Lee et al., 1994). The basic procedure is to determine lake levels and connecting channel flows by routing the simulated water supplies through the Great Lakes system with a hydrological response model (Hartmann, 1987; Quinn, 1978). In addition to net basin supplies, monthly diversions and consumptive uses data (International Great Lakes Diversions and Consumptive Uses Study Board, 1981) are also input to the model. GLERL's Hydrological Response Model consists of regulation plans, channel routing dynamics, and water balances, combined to estimate lake levels and connecting channel flows from water supplies to the lakes. Lake Superior is regulated by Plan 1977-A (International Lake Superior Board of Control, 1981, 1982) and Lake Ontario by Plan 1958-D (International St. Lawrence River Board of Control, 1963). The regulation plans were modified (Lee et al., 1994) and now have extreme condition operation rules. The modifications provided the robustness for the plans to handle the wider range of outflows expected during climate change and stochastic hydrological studies of the Great Lakes Basin than were used in the derivation of the Plans. In addition several minor modifications were made to allow the models to function under the extreme high and low lake levels and flows expected under severe transposed climates. Middle lake outflows are represented with stage-fall-discharge equations as functions of lake levels or of lake level differences between lakes. Flow retardation from ice and weeds are given by monthly median retardation values. Constant diversions are used for the Ogoki, Long Lac, and Chicago diversions and monthly means are used for Welland Canal diversions. Each lake storage, with all inflows and outflows, is described by mass continuity equations. The system of equations is solved numerically.

2.6 CLIMATE CHANGE AND REGIONAL SCENARIOS

2.6.1 Global climate simulations

Attention is increasingly being focused on forecasting/prediction of environmental changes which may result from changes in climate. For the Great Lakes region, rapid changes (i.e. warming) have the potential to alter physical (meteorological, hydrological, limnological) characteristics as well as water quality conditions and affect other socio-economic sectors. Potential changes in global/regional scale climate conditions are based on predictions from GCM's. Researchers have run GCM's of the earth's atmosphere to simulate climates for current conditions and for a doubling of global carbon dioxide levels ($2\times CO_2$). Currently, there are in excess of 10 GCM models being developed and tested. Examples of some GCM models whose outputs have been incorporated into large lake investigations include the GISS (Goddard Institute for Space Studies), GFDL (Geophysical Fluid Dynamics Laboratory), OSU (Oregon State University) and the Canada Climate Centre second generation model (CCCII). The following provides a synopsis of GCM considerations for application to regional scale analyses and case study approaches and implications on physical and water quality conditions in the Great Lakes based on changed climatic conditions.

2.6.2 Global circulation models (GCM's)

An example of GCM model structure and simulation results is given here for the Canada Climate Centre Version II GCM. The CCCII GCM has a vertical resolution of 10 layers from the surface to the 10mb level and horizontal resolution corresponding to a grid of 3.75°. latitude x 3.75° longitude (McFarlane et al., 1991). The model includes full diurnal and annual cycles and a complete representation of the hydrological cycle. Sophisticated computations are conducted for physical processes such as radiative transfer, cloud and cloud water content, surface energy balance and soil hydrology, and treatment of the ocean mixed layer and sea ice which includes a specification of oceanic heat transports that permit simulation of ocean surface temperature distribution and ice boundaries. CCCII GCM climate simulations are derived by running the model through several annual cycles until an equilibrium state is achieved. The results of the last few cycles are aggregated to form simulated monthly and seasonal climate statistics (McFarlane et al., 1991). Typical outputs at grid-point locations include air temperatures, winds, specific humidity, cloudiness, soil moisture, and radiation which represent some key meteorological inputs for regional scale modeling of climate impacts or potential responses to climate change conditions.

The results of enhanced CO_2 simulations made with several different GCM's including the CCC model are summarized and discussed in WMO/UNEP/IPCC (1990).

2.6.3 GCM limitations to regional-scale application

There are limitations to GCM model applications to regional scale analysis which have been noted by several investigators. For example, a comparison of monthly $2\times CO_2$ GCM outputs nearest to the central basin of Lake Erie (GISS, GFDL and OSU) is provided in Figure 2-27 for several key variables such as air temperature, wind speed and solar radiation (Schertzer and Sawchuk, 1990).

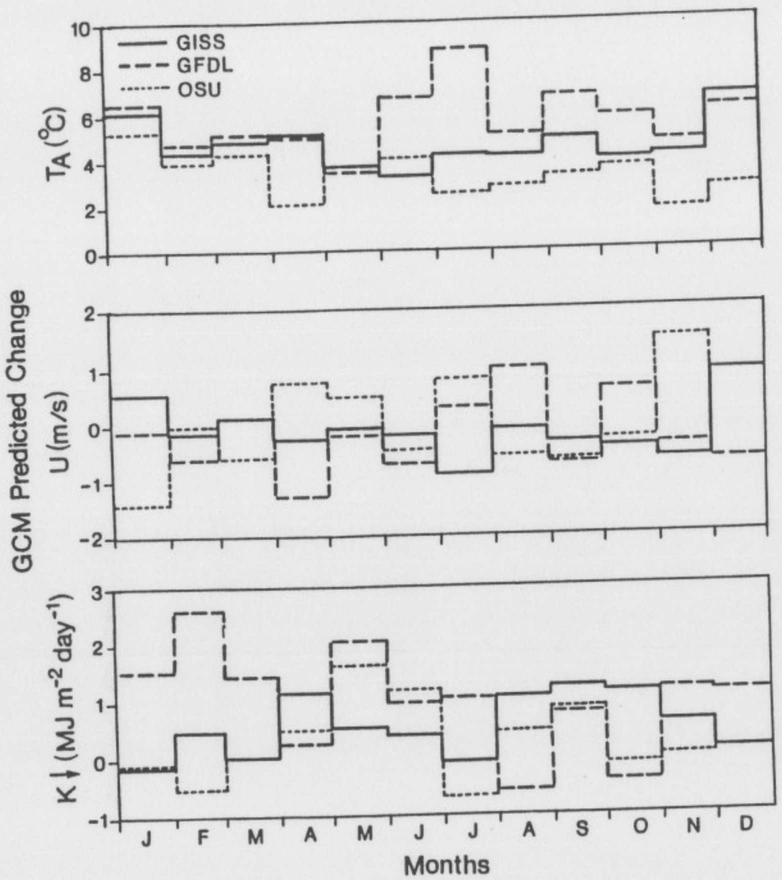

Figure 2-27. *Predicted changes in monthly air temperature T_A, wind speed U, and incoming solar radiation from the GISS, GFDL, OSU general circulation models for a doubling of atmospheric carbon dioxide concentrations for grid points nearest the lower Great Lakes (Schertzer and Sawchuk, 1990).*

There is broad qualitative agreement of seasonal warming, however, the range in predictions is large and in only a few instances do the models agree in both sign and magnitude of predicted change. Best agreement is for increases in air temperature. Predicted changes for wind speed and incoming solar radiation are frequently larger than the mean predicted value from the three models. Evaluation of the coefficients of variation for incoming solar radiation estimates (Schertzer and Sawchuk, 1990) showed rather large values because these estimates depend upon models of cloudiness and radiation transfer. Presently, cloud prediction is among the most uncertain elements of the radiative transfer in general circulation models (Simmons and Bengtsson, 1984; Henderson-Sellers, 1986). Predictions, for example, of global air temperatures may vary by as much as 2°C, depending upon the model. When GCM simulations of present regional climates are compared, the disagreements can be considerable, particularly for precipitation (Grotch and MacCracken, 1991).

Regional impact studies that have adopted GCM climate change scenarios have common uncertainties associated with the use of the outputs (Cohen, 1991). Examples of some of the common areas of uncertainty related to GCM outputs, scenarios, and regional models include a lack of a standard approach to derive regional climate change scenarios, no grid cells in GCM's classified as lakes or other sub-grid scale features, and GCM grid cells represent a weighted average of an area whereas station data represent a point within an area (Cohen, 1991). Unfortunately, the present-generation of GCM's specify all continental grid points as land. In GCM grid-point arrays for North America there are no nodes over the Great Lakes. Consequently, estimation of representative key "over-lake" meteorological input variables for climatological and hydrodynamic modeling requires additional procedures to reduce GCM scale outputs to sub-regional scale scenarios.

Model	Grid Cell Location		Grid Area (km^2)
	Latitude	Longitude	
GISS	7.83°	10.00°	674,144
GFDL	4.44°	7.50°	286,736
OSU	4.00°	5.00°	172,216
CCCII	3.75°	3.75°	121,086

Table 2-4. *Grid resolution of GCM models (Louie, 1991)*

2.6.4 Regional-scale climate scenarios

In general, GCM models do not provide meaningful simulations of sub-synoptic scale features; this is a significant limitation for formulation of climate change scenarios on the regional scale. Consequently, regional scale climate impact investigations on the Great Lakes are often forced to use inappropriately large spatial and temporal scales. For example, GLERL hydrological process models are defined over daily intervals and sub-basin areas averaging 4,300 km^2 while the GCM adjustments were made over monthly time intervals and grids of 7.83° latitude by 10° longitude (GISS), 4.44° by 7.5° (GFDL), 4° by 5° (OSU), and 3.75° by 3.75° (CCCII). Table 2-4 provides a comparison of areal grid resolution from four GCMs in the region of the Great Lakes (Louie, 1991).

Direct application of GCM predictions, regardless of uncertainties, to the estimation of over-lake meteorological fields can lead to errors in flux computations. First, lakes with large thermal inertia can substantially modify thermal and moisture properties of air flowing over them. As indicated previously, application of scaling relationships (Phillips and Irbe, 1978) to meteorological data collected at shoreline stations is commonly used to estimate representative over-lake values (e.g. Derecki 1975; Schertzer, 1987; Croley, 1989b). This approach requires knowledge of meteorological variability from several stations at or near the lake perimeter or about 1 station per 5,000 km^2, whereas GCM's provide predictions for only one grid point per

10^3 km^2 (see Table 2-4). In attempting to incorporate GCM outputs, many assumptions are invoked such as the presumption that over-water/over-land atmospheric relationships are unchanged and that other variables such as solar insolation at the top of and through the atmosphere on a clear day are unchanged under the changed climate, modified only by cloud cover changes (Croley, 1995).

Several objective approaches for transferring output from the GCM grid to the regional scale have been attempted and are recognized to have limitations. Simple averaging of GCM data ignores inter-dependencies in the various meteorological variables. Spatial averaging over a box centered on the GCM grid point tends to filter all variability that exists in the GCM output over that box. Interpolation between GCM grid point values has the tendency to preserve at least some of the spatial variability in the GCM output. However, little is known about the validity of various spatial interpolation schemes and, for highly variable spatial data, they may be inappropriate. Furthermore, much of the variability at the smallest resolvable scale of GCM's is, unfortunately, spurious (Croley, 1995). In either case, simple approaches involving ratios, averaging, or interpolation of GCM outputs do not address the spatial or temporal variability problem for a changed ($2\times CO_2$) climate. Essentially, the changed climate maintains the variability and seasonal patterns which are inherent in the base climate condition.

Developing representative sub-grid scale climate scenarios is a difficult problem. Dickinson (1987) suggests that since a single GCM grid box represents an area of 20,000 km^2, the results of GCM's are not sufficiently reliable for determining regional scale change, short-term variability, and extremes. Research into methods to apply GCM outputs to the regional scale are ongoing and include interpolation schemes, Limited Area Models (LAMS), application of regional climate models within the GCMs, application of synthetic data sets of extremes and nested mesoscale models.

Louie (1991) performed an interpolation from the original CCCII GCM grid to a finer grid (1°×1°) which was approximately 111 km × 77.5 km over the Great Lakes basin. This procedure provided 338 interpolated grid point values. Louie (1991) recognized that the procedure does not increase the spatial resolution of the GCM output, rather, it merely provides a consistent and objective means to estimate sub-grid values. The interpolation algorithm used was a simple bilinear scheme which uses the nearest four data points and weights them by their inverse distance squared. Although the procedure was selected to reduce the tendency for smoothing or enhancing the original GCM output fields, Croley (1995) determined that such problems exist in the use of the finer grid with the hydrology models.

Limited Area Models or LAMS (Dickinson et al., 1989) which project on a finer resolution may eventually reduce some of the uncertainties associated with the coarse grid resolution of current GCM's. Alternatively, application of regional climate models that are embedded in GCM's offer a possibility of determining scenarios at sub-grid levels. However, the massive computer requirements of this approach and their development have thus far limited the length of time series to 1 to 2 years in length (Bates et al., 1994). In another approach, long-term synthetic data sets developed for extremes in wet or dry conditions (Quinn and Changnon 1989) from historical records have limitations for climate change investigations at regional scales since climatic changes represented by the time series were not as large as many GCM's predict could happen in the future over the basin. Nested mesoscale numerical models developed for the Great Lakes basin (Bates et al., 1994) are not yet capable of generating multi-decadal series of conditions essential for climatic sensitivity studies (Croley, 1995).

2.7 CLIMATE CHANGE IMPACTS

Regional climate impact studies have been conducted in Canada, USA, and other parts of the world to assess potential impacts of climate change on lake hydrodynamics, water quality, and various economic sectors (Cohen, 1991). Several approaches have been attempted to assess probable responses. These include examining responses for years with anomalously warm conditions as an analogue of change to climate and impact studies which analyze responses by introducing change based on selected key meteorological variables from GCM steady-state and/or transient scenarios. In addition, transposition climates have been applied to regions in an attempt to evaluate the effect of climatic variability changes and the effect of extreme changes in climate. Preliminary impact estimates considered simple constant changes in air temperature or precipitation (Quinn and Croley, 1983; Cohen, 1986). The basic premise in many of the regional impact investigations is that regional models can link to GCM outputs to assess changes associated with climate change scenarios. A brief summary of examples of such case studies is presented here to illustrate some of the potential responses of large lake systems to climate changes.

2.7.1 Lake response to a warm year climate condition

Schertzer and Sawchuk (1990) examined the thermal structure of the lower Great Lakes for an anomalously warm year to infer the potential response for thermocline characteristics and anoxia occurrence in Lake Erie under observed extreme conditions as an alternative to applying GCM scenarios. The year 1983 was selected as an anomalously warm year. Primary elements of the lake thermal response to seasonal gains and losses in lake heat content include surface water temperature, ice cover, formation of the thermal bar and disappearance of the 4°C surface isotherm, thermocline depth and duration of thermal stratification period. Graphical representation of the warmer than average year of 1983 for surface water temperature, heat content and ice extent in comparison to long-term means is shown in Figure 2-28 for Lake Erie as an example.

Based on heat flux computations, it was determined that 1983 was characterized by large reductions in surface heat losses in winter and above average surface heat flux gains in summer. On an annual basis, the lower Great Lakes buffered large surface heat gains in summer months through losses in other months. Observations (Figure 2-28) for Lake Erie indicated higher surface water temperatures, significant reductions in extent and duration in ice cover, and earlier disappearance of the 4°C isotherm, signaling an earlier start to thermal stratification. In response to greater surface heating and low wind conditions, the thermocline formed higher in the water column, and stratification lasted longer than in other years (Figure 2-29). In the case of Lake Erie, these conditions contributed to slight hypolimnetic anoxia in the central basin in the latter half of September (Schertzer and Sawchuk, 1990).

2.7.2 Response of Lake Ontario to steady-state GCM scenario

Examples of predictions and limitations of simulations of thermal responses for Lake Michigan are discussed in detail in Section 3.4. We include here an example of a modeling exercise to determine the potential effects of a GCM climate scenario for

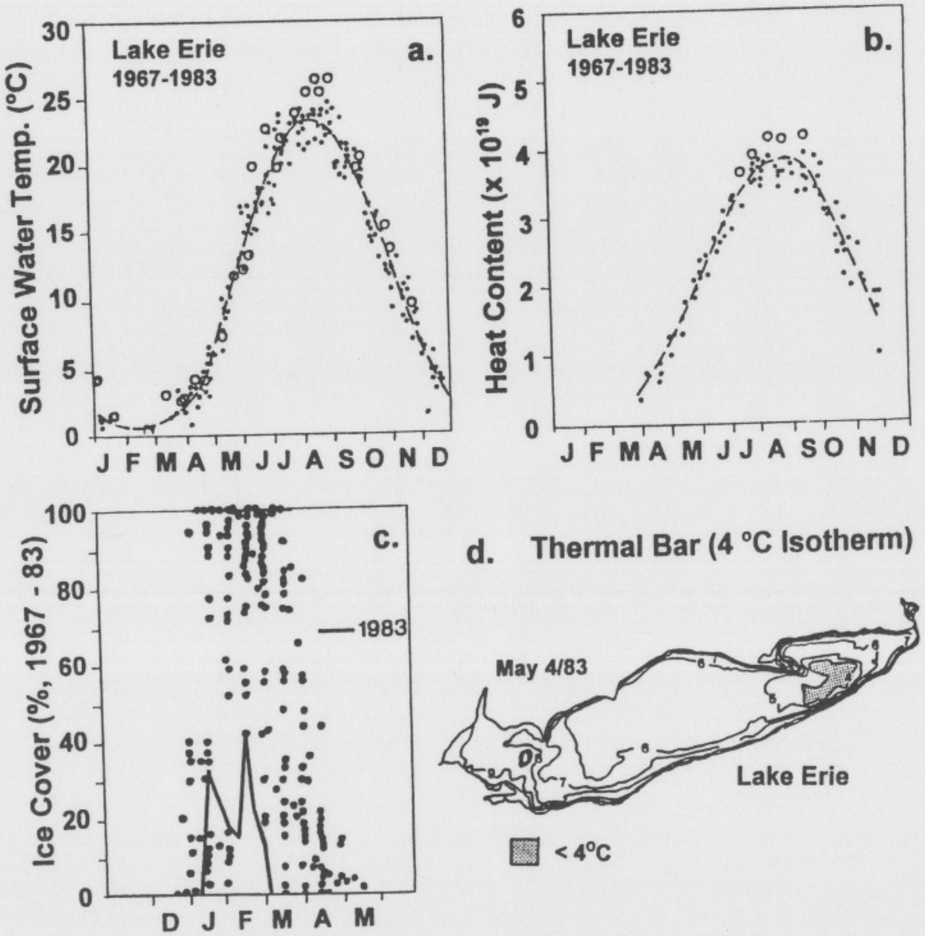

Figure 2-28. *An example of the observed thermal response of Lake Erie for an anomalously warm year 1983 compared to long-term observations (a) surface water temperature, (b) heat content and (c) ice cover and (d) disappearance of the thermal bar. (Schertzer and Sawchuk, 1990).*

Lake Ontario as an additional example of possible effects of climate change on large lake thermal regimes.

Boyce et al. (1993) examined the heat load on Lake Ontario and its thermal response under current conditions and for a hypothetical climate change scenario based on CCCII GCM output. Detailed measurements conducted in 1972 during the International Field Year for the Great Lakes (IFYGL 1981) were used as test data. A description of the model application to Lake Ontario is given by Boyce et al. (1993). Briefly, daily incoming solar radiation is based on detailed IFYGL measurements (Davies and Schertzer, 1974) and IFYGL surface meteorological data (i.e. air temperature, wind speed, relative humidity and cloudiness) were used to estimate latent and sensible heat transfers and incoming long-wave radiation. Daily hydrological inflow and outflow, temperature and salinity from major tributaries are specified. Outgoing long-wave radiation is computed from simulated water surface temperature. Simulation of the daily lake-wide averaged vertical temperature profile was done by the

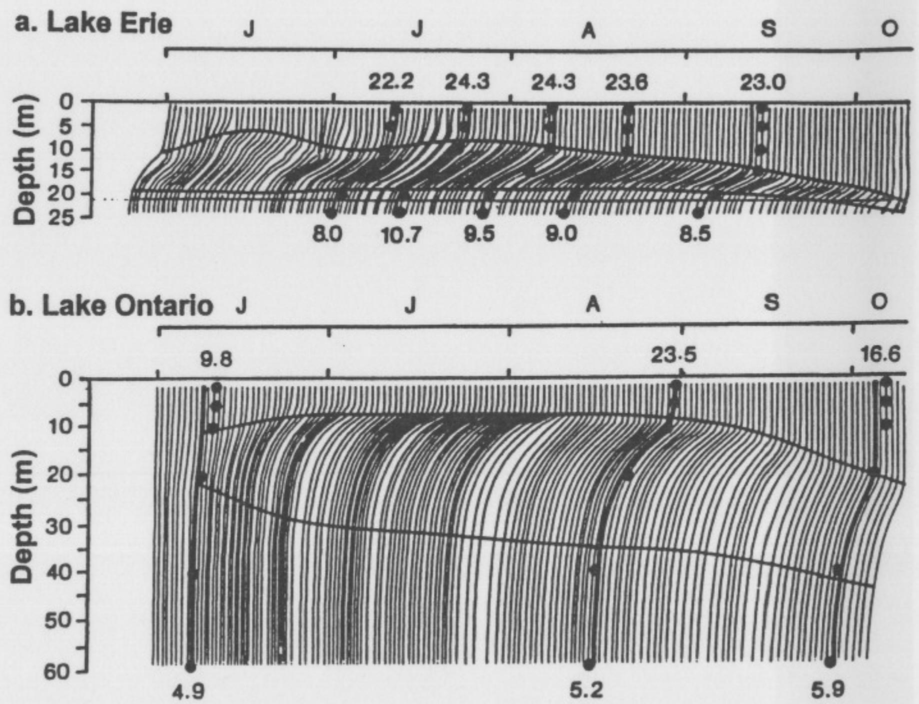

Figure 2-29. *Computed daily vertical temperature profiles (thin lines) for the central basin of (a) Lake Erie and (b) Lake Ontario in 1983. Heavy vertical lines represent temperature profiles evaluated from Lam and Schertzer's 1987 thermocline model for Lake Erie and Simon's (1980) thermocline model for Lake Ontario. Solid circles represent observed values determined from lake-wide ship cruises. Calculated depths of the base of the epilimnion and the top of the hypolimnion are illustrated with dark horizontal lines. Numerical values represent observed surface and bottom temperatures. (Schertzer and Sawchuk, 1990).*

the application of the one-dimensional Dynamic Reservoir Simulation Model (DYRESM) (Imberger and Patterson, 1981). DYRESM simulates the temperature profile by accounting for the creation, distribution, and dissipation of turbulent kinetic energy through wind stirring and surface heat flux.

CCCII predicted changes for key meteorological variables at the nearest grid point to Lake Ontario are shown in Table 2-5. As noted above, the nearest grid point of either GCM models do not include the Great Lakes in their parameterizations. Consequently, the assumption is made here that the GCM prediction for the grid square would be the same if Lake Ontario were included. The simulation proceeded by modifying input parameters from 1972 either by additive or multiplicative values (i.e. ratio $2\times CO_2/1\times CO_2$) which vary seasonally. The climate scenario used was to modify air temperature and cloudiness by adding the GCM differences to the field data while wind and atmospheric absolute humidity are modified by ratios. As indicated in Table 2-5, the CCCII scenario introduced large changes, primarily in the winter months.

Month	Air Temp. (°C)	Vapor Pressure (mb)	Wind Speed (m/s)	Cloud Cover (%)
January	6.77	1.43	1.06	0.0505
February	11.06	2.30	1.26	0.0287
March	8.67	1.84	1.22	0.0215
April	7.97	1.54	1.31	0.0429
May	4.85	1.38	0.98	-0.0125
June	4.91	1.36	0.97	-0.0031
July	4.77	1.31	0.96	-0.0162
August	4.76	1.26	0.86	-0.1079
September	4.26	1.24	0.85	-0.0384
October	3.75	1.26	0.94	-0.0261
November	2.07	1.12	0.89	-0.0830
December	2.63	1.06	0.93	0.0247

Table 2-5. *CCCII-predicted monthly mean changes for nearest Lake Ontario grid point (Boyce et al., 1993).*

Figure 2-30 shows horizontally averaged isotherms for Lake Ontario computed using the DYRESM model under (a) current climate conditions, and (b) a simulation based on the CCCII GCM. Figure 2-30b dramatically illustrates that in the climate change scenario, the simulated lake no longer experiences spring and fall convective overturn (4°C water at the surface). The stratified period is two months longer and maximum surface water temperatures are 4°C higher. Minimum summertime temperature throughout the lake is computed to be 6°C. A comparison of changes in selected surface heat flux components for the base run and the GCM scenario run indicated that both incoming and outgoing long-wave radiation fluxes are increased by similar amounts. From April through December, evaporative cooling of the lake is increased. In the winter months, January through March, the changes are extremely variable, both positive and negative. Sensible heat flux does not change greatly from May through November but the downward sensible heat flux increases markedly in the winter (January through April).

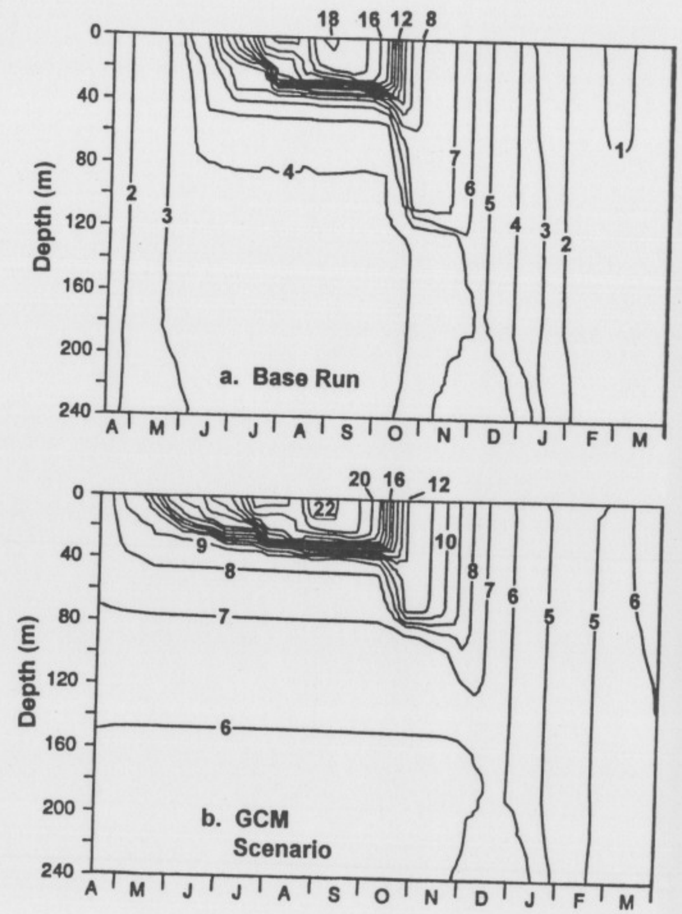

Figure 2-30. *Horizontally-averaged isotherms for Lake Ontario during IFYGL using the DYRESM model under (a) current (base) climatic conditions, and (b) using a $2\times CO_2$ CCCII GCM climate scenario. (based on Boyce et al., 1993)*

The sensitivity of predicted changes in thermal structure on Lake Ontario, was examined by arbitrarily varying air temperature and wind speed seasonally and annually over selected ranges (Boyce et al., 1993). By increasing only air temperature by 4°C annually, the simulated results indicate that epilimnion water temperature increases with the largest increase occurring in the April-September period. Temperature gradients across the thermocline are increased and the stratified period is lengthened. Complete overturn, which occurs at the end of December in the base run, is delayed by approximately a month.

Wind speed was modified to allow for a 40% annual increase and decrease to assess probable changes in thermal structure. A 40% decrease in annual wind speed resulted in the formation of a shallower mixed layer and intensification of thermal gradients across the thermocline. Alternatively, a 40% increase in annual wind speed resulted in a substantial mixed layer deepening by about 10m.

2.7.3 Climate impacts on hydrology (steady-state / transient models)

The U.S. Environmental Protection Agency (USEPA) and the International Joint Commission (IJC) climate change studies conducted by the Great Lakes Environmental Research Laboratory (GLERL) investigated mean responses in hydrological variables. Highlights of these investigations (USEPA, 1989; Croley 1992b) with respect to lake hydrology are presented here.

2.7.3.1 GLERL-EPA 2×CO_2 climate impacts

As part of an EPA study, the Great Lakes Environmental Research Laboratory (GLERL) assessed steady-state and transient changes in Great Lakes hydrology consequent with simulated 2×CO_2 atmospheric scenarios from three GCMs (Croley, 1990; Hartmann, 1990; USEPA, 1989). EPA required that GLERL first simulate 30 years of "present" Great Lakes hydrology by using historical daily data with present diversions and channel conditions. GLERL arbitrarily set initial conditions but used an initialization period to allow their models (described earlier in subsection 2.5.4) to converge to conditions initial to the simulation. GLERL repeated their simulation, with initial conditions set equal to the averages over the simulation period, until these averages were unchanging. This facilitated investigation of "steady-state" conditions. The next step was to conduct simulations with adjusted data sets.

EPA obtained output from atmospheric GCM simulations, representing both "present" and 2×CO_2 steady-state conditions, from GISS, GFDL, and OSU. They supplied monthly adjustments of "present" to 2×CO_2 for each meteorological variable. GLERL applied them to daily historical data sets to estimate 33-year sequences of atmospheric conditions associated with the 2×CO_2 scenarios. This method unfortunately keeps spatial and temporal (inter-annual, seasonal, and daily) variability the same in the adjusted data sets as in the historical base period. GLERL then used the 2×CO_2 scenarios in hydrology impact model simulations similar to those for the base case scenario. They interpreted differences between the 2×CO_2 scenario and the base case scenario as resulting from the changed climate. They observed that the three scenarios changed precipitation little but snow-melt and runoff were greatly decreased, evapotranspiration and lake evaporation were greatly increased, and net basin supplies to the lakes and lake levels were decreased. The scenario derived from the GFDL GCM was the most extreme with evaporation 44% higher than the base case and net basin supply less than 50% of the base case. Results for the entire Great Lakes basin were assembled by integrating all of the sub-basin and lake simulations; they are summarized in Table 2-6 for steady-state studies.

Scenario	Overland Precip.	Evapo-transpir.	Basin Runoff	Over-lake Precip.	Over-lake Evap.	Net Basin Supply
Base[a] (m^3 s^{-1})	13855	7814	6206	6554	4958	7803
GISS[b]	2 %	21 %	-24 %	4 %	26 %	-37 %
GFDL[c]	1 %	19 %	-23 %	0 %	44 %	-51 %
OSU[d]	6 %	19 %	-11 %	6 %	26 %	-23 %
CCC[e]	-2 %	22 %	-32 %	0 %	32 %	-46 %
6°S × 10°W[f]	+ 6 %	+31 %	-25 %	+ 3 %	+49 %	-48 %
6°S × 0°W[f]	+24 %	+43 %	- 1 %	+25 %	+33 %	- 1 %
10°S × 11°W[f]	+17%	+ 48 %	-21 %	+13 %	+75 %	-54 %
10°S × 5°W[f]	+45 %	+ 78 %	+ 2 %	+45 %	+69 %	- 5 %

[a]Base Climate (present conditions) from Transposition Study (Croley et al., 1996).
[b]Goddard Institute for Space Studies GCM (Croley, 1990).
[c]Geophysical Fluid Dynamics Laboratory GCM (Croley, 1990).
[d]Oregon State University GCM (Croley, 1990).
[e]Canadian Climate Centre (Croley, 1993a).
[f]Transposed Climates from the Southwestern US (Croley et al., 1996).

Table 2-6. *Average annual steady-state Great Lakes basin hydrology summary depicting current (base) climate conditions and potential changes using GCM and transposition climate scenarios.*

GLERL simulated one transient case over the period 1981-2060 with historical data from 1951-80, repeated three times, by using initial conditions of 1 January 1981 for the first 30 years; the second used the end-of-run conditions from the first simulation as initial conditions and the third used end-of-run conditions from the second. The EPA supplied GISS GCM transient scenario A, consisting of 9 sets (for each decade from 1970-9 through 2050-9) of monthly ratios and differences of meteorology between the "present" and the transient "future" representing an increasing atmospheric-CO_2-content over the period 1971-2059. GLERL interpolated to obtain adjustments for each month of 1981-2059 and applied them in three simulations as for the base case. GLERL used a differencing approach to discern the 2×CO_2 signal from the historical variations in the adjusted data sets, comparing values 30 years apart to eliminate the (repetitive) historical variations. GLERL then interpreted differences between the transient scenario and the base case scenario as resulting from the changing climate, summarized in Table 2-7. Other EPA studies included partial assessments of large-lake heat storage associated with climate change on Lakes Michigan (McCormick, 1989) and Erie (Blumberg and DiToro, 1989).

Hydrological Variable	Units	Basin					
		Sup.	Mic.	Hur.	StC.	Eri.	Ont.
Basin Air Temperature	(°C/dec):	+0.5	+0.6	+0.6	+0.6	+0.6	+0.6
Annual Basin Precip.[a]	(mm/dec):	+29	+7	+2	-0	+3	+1
Annual Basin Evap.[a]	(mm/dec):	+25	+14	+12	+15	+16	+16
Annual Basin Runoff[a]	(mm/dec):	+4	-7	-9	-15	-15	-14
Snow Pack[a]	(mm/dec):	-4	-1	-4	-1	-0.7	-2
Soil Moisture[a]	(mm/dec):	+0.5	-0.8	-1	-0.4	-0.4	-0.6
Groundwater[a]	(mm/dec):	+2	-1	-0.3	-0.4	-0.4	-0.3
Total Basin Moisture[a]	(mm/dec):	-0.1	-4	-5	-2	-2	-4
Lake Air Temperature	(°C/dec):	+0.7	+0.6	+0.6	+0.6	+0.6	+0.6
Lake Humidity	(mb/dec):	+0.6	+0.5	+0.5	+0.7	+0.7	+0.6
Lake Cloud Cover	(%/dec):	+0.1	-0.2	-0.5	-0.4	-0.5	-0.5
Lake Wind Speed	(m/s/dec):	+0.0	+0.0	+0.0	+0.0	+0.0	-0.0
Surface Temperature	(°C/dec):	+0.7	+0.5	+0.6	+0.5	+0.6	+0.6
Annual Lake Evap.[b]	(mm/dec):	+18	+19	+22	+38	+40	+24
Annual N.B.S.[b]	(mm/dec):	+17	-27	-41	-245	-75	-75
Annual Net Outflow[b]	(mm/dec):	+20	-31	-31	-241	-70	-57[c]
Lake Level	(mm/dec):	-13	-59	-59	-64	-66	-93[c]

[a]Expressed as a depth over the basin., [b]Expressed as a depth over the lake.
[c]Computed over first 7 decades since Ontario regulation plan fails in eighth.

Table 2-7. *GISS transient climate changes impacts summary (Croley, 1995).*

2.7.3.2 GLERL-IJC $2\times CO_2$ climate impacts

The EPA studies, in part, and the high water levels of the mid 1980s prompted the International Joint Commission (IJC) to reassess climate change impacts on Great Lakes hydrology and lake thermal structure. The IJC study looked in less detail but more breadth at large-lake thermodynamics in that while only lake-wide effects were considered, all lakes were assessed. GLERL adapted the EPA study methodology for the IJC studies (Croley, 1992b) to consider $2\times CO_2$ GCM scenarios supplied by the Canadian Climate Centre (CCC) for the period 1948-88. GLERL's procedure to estimate "steady-state" suggested, for a few sub-basins, very different initial groundwater storages than were used in model calibrations. Since there is little confidence in estimates of very large groundwater half-lives on these sub-basins with only 10 to 20 years in calibrations, those initial values used in calibrations were also used in the simulations for those sub-basins.

Average monthly meteorological outputs were supplied for each month of the year over a 1° latitude by 1° longitude grid (Louie, 1991) by the CCC as resulting from their second-generation GCM; see McFarlane (1991). GLERL computed $2\times CO_2$ monthly adjustments at each location, used them with historical data to estimate the $2\times CO_2$ 41-year sequences (1948-88) for each Great Lake basin, and then used the $2\times CO_2$ scenario in simulations similar to the base case as before. This scenario proved similar to the earlier GFDL-based scenario in that net basin supplies were reduced to almost 50% of the base case. However, the CCC-based scenario reduced runoff more and evaporation less than the GFDL-based scenario. Results for the entire Great Lakes basin again are summarized in Table 2-6 to enable comparison with the EPA steady-state study results.

2.7.4 Great Lakes response to transposed climates

GLERL integrated their hydrological process models into a system to estimate lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1992b; Croley and Hartmann, 1987; Croley and Lee, 1993). This system was used to simulate the Great Lakes hydrology for historical meteorology and several transposed scenarios.

For many Great Lakes issues, the impact of climatic variability is an important consideration. From a hydrological perspective, variability includes consideration of such factors as shifts in the daily, seasonal, inter-annual, and multi-year climate variability on lake net supply behavior, as well as related changes in mean supplies. The difficulty here is attempting to assemble a long-term data base capable of assessing variability and sensitivity of fluctuations in the hydrological system of the Great Lakes. Transposing climates from one region to another offers a viable alternative to assess response and sensitivities to an altered climatic condition.

2.7.4.1 Climate transposition scenarios

Climate transposition is an empirical technique to impose meteorological time series from one location to another in order to examine responses based on a new range of temporal variability and frequency and magnitude of extremes. Transposition of actual climates incorporates natural changes in variability within the existing climate; this is not true for GCM-generated corrections applied to existing historical data in many other hydrological impact assessment studies. Further, development of detailed scenarios by transposing climates through station relocation avoids the problem of properly incorporating physically plausible climatic characteristics on the existing network of 2000 stations over a climatological time frame of say 40 years.

GCM's predict that continuing increases in atmospheric trace gas concentrations will result in warmer conditions, comparable to climates south of the Great Lakes. Some GCM's also predict drier conditions, comparable to climates to the west of the Great Lakes. Therefore, the future climate of the Great Lakes may be similar (at least in terms of annual means and other very general features) to the present climate of regions to the south and west of the Great Lakes. Transposed scenarios were selected to represent analogues of "future" climatic conditions. Current climatic conditions to the south and west of the Great Lakes were used to develop climate scenarios under the assumption that future changes in the basin climate may approximate latitudinal and/or longitudinal climatic shifts (Croley et al., 1996). Scenario 1 (warm and dry) corresponds to warmer temperatures and mixed precipitation changes. It represents a movement of the Great Lakes Basin 6°S and 10°W resulting in a climate similar to the central Great Plains, middle Mississippi, and lower Ohio valleys. Scenario 2 (warm and wet) is a simple shift 6°S and corresponds to warmer temperatures and increases in precipitation amounts over the entire basin. Scenario 3 (very warm and dry) corresponds to very high temperatures and mixed precipitation changes. It is a shift 10°S and 11°W and, while generally wetter over much of the basin, is drier in the western part of the basin. Scenario 4 (very warm and wet) corresponds to very high temperatures and large increases in precipitation over the entire basin. It is a shift 10°S and 5°W.

The relative spatial relationships of the geography of the Great Lakes were preserved with the outline of the basin laid over the existing climate network. Station data from the new locations were translated to the Great Lakes to derive new station networks and basin climates by using Thiessen weightings over the 121 sub-basins and seven lake surfaces of the Great Lakes Basin. Scenarios 1 and 2 correspond roughly to the upper range of GCM predictions for temperature for the Great Lakes basin (IPCC, 1992) while Scenarios 3 and 4 went beyond the range of current GCM predictions for a doubling of atmospheric trace gas concentrations in order to determine how the Great Lakes would respond to a major climatic shock.

Scenario climatology was examined to determine the representativeness of transposed climate gradients for climate change analysis (Croley et al., 1996). With respect to mean temperatures, north-south gradients were similar to current Great Lakes basin conditions. Precipitation patterns were more complex with generally higher precipitation in all scenarios, however, differences in seasonal, and west-east gradients place constraints on the precipitation applicability. Total annual snowfall is significantly decreased in the transposed climates. Relative humidity changes were found to be scenario dependent (i.e. scenarios 2 and 4 had little change while scenarios 1 and 3 have decreases especially in the western part of the basin). The transposed scenarios all projected decreases in cloudiness which implies increases in solar radiation amounts. For the most part, wind speed gradients in the transposed scenarios were similar to existing conditions. Average water vapor pressure deficits generally increase in all scenarios.

Croley et al. (1996) provide details of the physical process models applied in this analysis as well as calibration procedures, validity, and applicability (also briefly described earlier in subsection 2.5.4). Base case and climate transposition scenarios were analyzed with the models to assess the response of the Great Lakes basin and lakes for selected meteorological variables, basin hydrology, over-lake meteorology, lake heat balance, lake thermal structure, and lake water balance as well as an examination of hydrological sensitivities.

Although the climate transposition approach has limitations in that several climatic variables influencing the lake (i.e. temperature and precipitation) are interrelated, a wide range of potential basin and lake responses together with estimates of variability can be extrapolated from the results. Brief numerical results are summarized in Table 2-6 for ready comparison with the EPA and IJC studies. Generalized observations follow.

2.7.4.2 Lake evaporation increases

All scenarios produced significant increases in lake evaporation. From an energy standpoint, the energy for increased evaporation is derived from four sources. Most important is the decrease in cloud cover which results in increased incoming solar radiation and, on average, accounts for about half of the evaporation increase. The second source is increased downward long-wave radiation emitted by the atmosphere, a result of the higher temperatures; this accounts for 10-15% of the effect. These increased radiative sources result in a greater accumulation of heat in the lakes. A third important factor is a change in the partitioning of energy between sensible and latent heat flux. As a result of the saturation vapor pressure-temperature relationship, the ratio of sensible to latent heat flux (Bowen ratio) decreases in all scenarios and accounts for about one-third of the effect. A fourth factor is a decrease in lake ice cover. Because of the higher temperatures, nearly all lakes remain ice-free throughout the winter. Thus, the average albedo during the winter and early spring months is lower, increasing the amount of solar radiation absorbed by the lake. However, this makes a minor (1-2%) contribution.

The increase in downward long-wave radiation, the change in the partitioning between sensible and latent heat flux, and the decrease in lake ice cover result from fundamental physical principles, and they will almost certainly be a feature of any climatic state warmer than current conditions in the basin. There will thus be a considerable positive pressure on lake evaporation. However, the changes in cloud cover in these scenarios may not be realized in a future warmer climate. Thus the increases in lake evaporation in these scenarios would be smaller if cloud cover does not decrease.

Another interesting aspect of lake evaporation is that it is highly event oriented. A large proportion of evaporation occurs during Arctic cold air outbreaks in the cold season. In all scenarios for all lakes, the relative contribution of these events to total lake evaporation increases. Although the increase may be unique to these specific scenarios, these events are also important in the current climate. This indicates that accurate future estimates of lake evaporation will require accurate estimates of the number and severity of cold air outbreaks.

2.7.4.3 Soil moisture and runoff reductions

Many scenarios result in lower soil moisture and reduced runoff despite higher precipitation. As a result of the higher temperatures and longer growing seasons, the four scenarios produced a more vigorous overland hydrological cycle. Total annual evapotranspiration from the ground and the vegetation increases in all scenarios. Also, higher temperatures significantly reduced total snowfall. In the current climate, the spring snow-melt runoff season is very important to the total lake hydrology. In the four scenarios, the snow-melt season is shorter and less significant. The above results are likely to be a feature of any warmer climate. This means that in a warmer climate, greater precipitation is required to maintain runoff at present levels.

Figures 2-31a and 2-31b illustrate the interrelationships between air temperature and runoff/precipitation ratios and lake evaporation changes for the Lake Superior basin, the most sensitive basin for this study. The figures also include data from the previously mentioned EPA and IJC studies. The relationship among these variables is amazingly linear. This suggests that, for very broad annual impact assessments over an entire Great Lake basin, very simple relationships can be used in the investigations. Figure 2-31a shows that substantial increases in precipitation would be required to maintain runoff equivalent to the present climate.

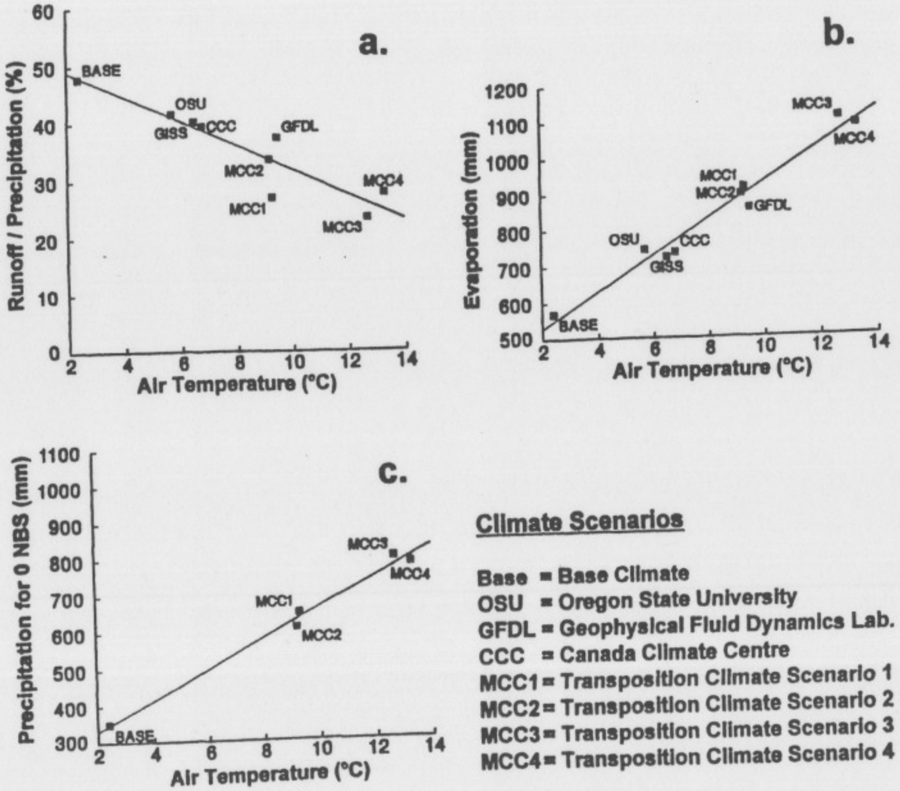

Figure 2-31. *Variation of selected components with temperature under changed climates (a) Average annual Lake Superior basin effective-runoff variation with temperature under changed climates, (b) Average annual Lake Superior evaporation variation with temperature under changed climates, (c) Average annual Lake Superior precipitation required for zero NBS variation with temperature under changed climates (Croley et al., 1996).*

2.7.4.4 Net basin supply decreases

Warmer climates result in large negative pressures on net basin water supply. Net basin supply (NBS) is comprised of the sum of over-lake precipitation and surface runoff into the lake, minus lake evaporation. The previous two findings have indicated that warmer climates will likely lead to increases in lake evaporation and decreases in runoff. Thus, significantly greater precipitation is required to maintain NBS at current levels of the Great Lakes. There is a very coherent relationship among NBS changes and the changes in temperature and precipitation. The relationships in Figs. 2-31a and 2-31b can be combined to build a relationship between precipitation required to maintain annual net basin supply totals and air temperature in a changed climate (see Figure 2-31c). Figure 2-31c suggests that for temperature changes of 5 to 6°C, precipitation increases of 20 to 30% may be required to maintain Lake Superior NBS at today's levels. If annual mean temperatures were to increase with no compensating increases in precipitation, it is highly likely that NBS levels would fall significantly.

2.7.4.5 Net basin supply variability increases

These scenarios produce much higher variability in NBS. Inter-annual variability in NBS, expressed as an over-lake depth, ranges from 140 to 300 mm under current climate changes. These scenarios produce average increases of about 60% in warm scenarios 1 and 2 and about 90% in very warm scenarios 3 and 4. These increases are primarily due to increases in precipitation variability in these scenarios. The Great Lakes currently experience lower precipitation variability than that of locations to the west and south of the basin. Thus, all scenarios have increased precipitation variability. Kunkel et al. (1993) have pointed out that a major contributor to inter-annual precipitation variability in the Great Lakes region are infrequent large multi-day precipitation events. Thus, accurate estimates of precipitation variability expected in future climates will require an accurate simulation of the frequency and magnitude of these infrequent large events.

The changes in lake evaporation variability are rather small (< 20%) and are thus not a major factor in changing NBS variability. However, in some scenarios, the simulated increases in runoff variability are caused partially by higher variability in evapotranspiration. This is a result of the longer growing season which results in a greater exposure to soil moisture stress. However, this is a minor factor compared to the contribution of precipitation variability.

2.7.4.6 Reduced turnover frequency

Warmer climates can result in reduced frequency of buoyancy-driven water column turnovers. In many of these scenarios, lake surface water temperatures often do not fall to 3.98°C (the temperature of the maximum density of water) during the colder half-year. This was true also for the earlier EPA and IJC studies described herein. As a result, buoyancy-driven vertical turnovers of the water column change from a frequency of two times/year to once per year (see Section 3.4). Since this is related to a fundamental physical property of fresh water, it is highly likely that this will occur in any future climate that is sufficiently warm. This could result in significant environmental impacts since these turnovers are important for nutrient distribution, oxygenation of lake water (see Sections 7.7.4 and 7.8.2), and so forth.

2.7.4.7 Lake effects

Lake effects on regional climate have negligible hydrological effects. GLERL utilized existing spatial and quantitative measures of lake effects on various climate conditions to modify climate data for one of the scenarios. They tested lake effects on basin hydrology by calculating outcomes with and without lake effects present. The differences on runoff and lake levels for the various Great Lake basins were negligible. They did not attempt a modeling investigation to ascertain how much lake effects might change in warm-wet scenarios like 3 or 4, but the lack of differences suggest that huge changes in lake effects would be required to significantly alter the hydrological results.

2.7.4.8 Lake levels and outflows

Great Lakes water levels are lower and more variable under the transposed climates. A warmer climate over the Great Lakes basin, whether wetter or dryer, would have major impacts on Great Lakes water levels and flows in the connecting channels. Superior is the most sensitive lake in the system when looking at climate transposition. Under scenarios 1 and 3, the lake would have negative water supplies for all or part of the time. Under scenario 3 Lake Superior would become a terminal lake. In other scenarios, both record high and record low lake levels would be achieved. In addition most scenarios indicated greater variability of lake levels, both seasonal and inter-annual, than exist in the present regime.

From an adaptive viewpoint the individual lakes, with the exception of Lake Superior during negative water supply scenarios, could be regulated to maintain water levels at about the long term average. However, there is no way to maintain the connecting channel flows about their long means. There would also likely be a major effort to divert additional water into Lake Superior from the Hudson Bay watershed under low or negative water supply scenarios.

2.8 POTENTIAL IMPACTS OF GLOBAL CLIMATE CHANGE ON LAKE HYDROTHERMAL DYNAMICS IN OTHER CLIMATIC REGIMES

The potential response of freshwater lakes to changed atmospheric conditions has also been assessed for other lake systems in addition to the Great Lakes (i.e. Stefan and Ford, 1975; Hondso and Stephan, 1991; Kundzewicz and Somlyody, 1993). For large lake systems, changed climate and hydrological conditions result in consequences for such concerns as lake stratification patterns, vertical mixing, water quality, fisheries, and lake management in general.

Meyer et al. (1994) addressed the possible impact of climatic changes on the thermal regime of lakes for different geographic zones of the globe. Ice cover and convective overturn events were primary indicators for assessing lake responses to climatic change. The assessment considered the thermal responses of hypothetical lakes for shallow, intermediate, and deep lakes as well as evaluation of the responses of nine real lakes from sensitive regions. Future climate scenarios were based on the GFDL GCM under the assumption of $2\times CO_2$ atmospheric concentration. Regional sensitivity analyses of lake stratification on air temperature change were performed for all latitudes for every 5 degree of latitude.

2.8.1 Hypothetical lake responses to climate change

Hypothetical lake morphology was selected to characterize shallow, intermediate, and deep lakes. The surface area of the hypothetical lakes was limited to 100 km^2 and a linear dependency was assumed between lake depth and the area of horizontal cross-section. Five values for lake depths were selected, namely 10, 20, 50, 75, and 150 m. Latitudes were changed systematically by 5 degrees. Initial conditions for the simulation were depth of the epilimnion ($h(t = 0) = 0.01$ m); hypolimnion temperature (T_H = 8°C); and epilimnion temperature (T_E = 8.05°C) (Meyer et al., 1993). A vertically-one-dimensional model of deep stratified reservoirs with full mixing in horizontal layers was used to simulate the vertical temperature structure. The model selected was based on the Institute for Water and Environmental Problems Model (IWEP) from the Siberian Branch of the Russian Academy of Sciences. A description of the heat source (surface heat flux sub-model) and diffusion coefficient is provided by Meyer et al. (1994). Essentially, the turbulent portion of the effective diffusion coefficient is calculated as a function of turbulent kinetic energy and dissipation rate. Compared to other simulation models tested, the IWEP model had a shorter simulation time and could simulate ice cover and overturns.

The results of this analysis suggested that the sensitivity of lake stratification to changes in air temperature greatly increases in transition zones such as the subtropics (30°-45° N/S latitude), where lakes can change from warm monomictic to dimictic, and the sub-polar zone (65°-80° N/S latitude), where lakes can change from dimictic to cold monomictic. In subtropical regions shallow and intermediate lakes, stratification was slightly sensitive to changes in air temperature while significant sensitivity was found in deep lakes. In temperature and polar regions, sensitivity of lake stratification to changes in air temperature was important for all lake depths. Turn-over characteristics of subtropical and sub-polar lakes were significantly affected by warmer atmospheric temperatures as turnover began earlier than under current conditions and the well mixed layer duration was longer. Cooler atmospheric temperatures were found to delay the onset of lake turnover and reduce the period of well mixed conditions.

In sub-polar and polar regions, the duration of ice cover was most sensitive to changes in air temperature regardless of depth. It was found that changes in the duration of overturn period by as much as 10 days were common for a change of air temperature of 5°C. For eutrophic lakes, such changes are significant since mixing events are critical for alleviating water quality problems resulting from thermal stratification such as hypolimnetic anoxia. This investigation suggested that the effect of changing climate is equivalent to the corresponding change in geographic location (approximately one latitude degree per one degree Celsius of air temperature) (Meyer et al., 1993).

2.8.2 Response of lakes in sensitive regions to climatic change

Analyses for hypothetical lake responses to climate change indicated that changes were highly sensitive near transition latitudes. Transition regions were found to be sensitive belts of the globe with latitudes ranging from 30° to 45° N/S and 65° to 80° N/S. The lakes within the transition zones included Shasta Lake (California, USA), Great Bear Lake (NWT, Canada), Lake Seneca (New York, USA), Ezequiel R. Reservoir (Neuquen Provine, Argentina), Lake Geneva (Switzerland and France), Lake Maggiore (Italy and Switzerland), Changshou-Hu Reservoir (Sichuan Province, People's Republic of China), Lake Ladoga (Russia) and Lake Biwa-Ko (Shiga Prefecture, Japan). Historical data were used to simulate current conditions while climate changed conditions were based on GFDL GCM results.

The study revealed that in the warmer latitudes, periods of stratification may be enhanced. In colder higher latitudes, the frequency of overturn is likely to increase and there is the potential for sub-polar lakes to change from cold monomictic to dimictic. Ice formation in sub-polar regions have the potential to be reduced or totally suppressed. Climate change (increased air temperature) has the potential to increase water temperature significantly enough to induce changes in water quality and lake biota and changes in the local climate near the lake periphery.

2.9 SUMMARY

The Great Lakes basin and the Great Lakes are a dominant feature on the North American continent. Climate conditions (weather) vary considerably across the basin. Gradients in key meteorological variables are oriented primarily north-south, however, the lakes themselves also influence the intensity of west-east gradients. Mesoscale investigations have clearly shown lake effect influences on the lake periphery.

The climate, hydrology, and limnological responses of the Great Lakes basin are variable. There are seasonal cycles in meteorological variables which have a significant impact on the seasonal cycle of hydrological and limnological components. Long-term data indicate that the climate impacts both hydrological as well as lake thermal conditions. These have extended influences on other hydrodynamic processes as well as water quality conditions especially in shallower regimes such as Lake Erie.

Examples of several methodologies for assessment of potential responses of the Great Lakes to climatic change (warming) were briefly presented. These included analysis of anomalously warm years, climate impact studies using GCM steady-state or transient outputs, climate transposition approaches, and analysis of hypothetical lake responses to latitudinal changes in climate characteristics. It was shown that GCMs have limitations to their application on a regional scale and all methods have advantages and disadvantages which constrain their application. Analysis of warm year climate responses provide some indication as to the direction of responses for selected components. Application of GCM outputs to regional studies tend to provide information on the possible range of physical lake responses. Climate transposition analyses provide some measure of the variability which could be expected by translation of an existing warmer climate to the Great Lakes basin. Hypothetical lake responses to climate at varying latitudes indicated that there are sensitive climatic transition regions where lake hydrodynamic and thermodynamic responses may be more pronounced.

The case studies that were discussed provided information on potential changes to basin hydrology and lake thermal conditions. Since the Great Lakes basin is so vast, and the Great Lakes themselves differ in physical dimensions and heat storage capacity, responses to climate or climatic scenarios is somewhat dependent on location. In general, a warmer climate could be expected to increase both evapotranspiration and evaporation from the lakes. Changes in precipitation could result in decreased soil moisture and runoff especially impacting on reducing spring snow-melt runoff. The direction of change in hydrological components is such that the Great Lakes basin may see a reduction in net basin supplies which will adversely affect flows in connecting channels. One transposition scenario suggested that such reductions could result in Lake Superior becoming a terminal lake which would have serious impact on downstream lakes. With respect to lake hydrodynamics, warmer conditions with reduced wind mixing can affect thermal stratification characteristics of the Great Lakes. From a water quality perspective, changes in a shallower regime such as Lake Erie central basin can be more critical. Under anomalously warm conditions, with reduced wind speeds, investigations indicated lower ice cover extent, warmer water temperatures, and an increase in the lake thermal stratification period with a shallower upper mixed layer in Lake Erie. Application of GCM steady-state outputs to Lake Ontario indicated that it was possible for the lake to no longer experience spring and fall convective overturn (4°C water at the surface). In one simulation, the stratified period was two months longer and the maximum surface water temperatures were 4°C higher with an increased minimum summertime temperature throughout the lake. Climate transposition experiments reinforced the previous investigations, concluding that significantly warmer climate reduces the frequency of buoyancy-driven water column turnovers from a frequency of two times/year to once per year. Significant hydrodynamic and environmental impacts are anticipated with warmer conditions and especially with reduced turnover frequency. With respect to water quality considerations, turnovers are important for such processes as vertical mixing of nutrients and oxygenation of lake water (Lam et al., 1987).

Examination of hypothetical lakes over a range of latitudinal climates tended to reinforce some of the potential climatic change impacts on the thermal regimes determined for the Great Lakes. In addition, the analysis incorporating hypothetical lakes demonstrated that there are latitudinal zones in which lakes of varying sized may be particularly sensitive to climate changes.

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